to the Bay of Fundy, on the Nova Scotian side; two were from the New Brunswick side of the Bay of Fundy; and three were from the southern side of Grand Manan; totaling 9 per cent of the number set out. These drifts so closely reproduce (in their regional distribution) the recoveries of bottles set afloat farther out along this same line the year before (line A; fig. 181) and off Cape Ann in 1923 (fig. 176) that most of them, no doubt, followed a uniform route, at least in their journey northward toward the mouth of the Bay of Fundy.

It is evident that most of the bottles from line E moved offshore from the time of their release, otherwise more strandings would have been reported from the coast line to the southward. However, the fact that one of them was recovered at Nantucket, a second on the outer shore of Cape Cod, and a third at Beachwood, Me. (Nos. 1702, 1712, and 1731; intervals, respectively, of 138, 121, and 34 days), makes it probable that they followed a southeasterly course at first. Negative evidence to the same effect results from the fact that only two bottles from this line were found anywhere along the coast of Maine between Seguin Island and the Bay of Fundy, contrasting with the considerable number of recoveries from Nova Scotia. Had the set been eastward along the coast of Maine, such as would be represented by a straight line between the points of release and recovery, a considerable number of recoveries might have been expected along that 140-mile sector, where the tide draws strongly into the numerous bays, bringing in large amounts of drift of all kinds. It is fair to assume, also, that the route across the gulf was about as long for series E as for the Cape Ann series, because three of the latter were reported from Nova Scotia after intervals as brief as any from the northern lines; one, namely from Yarmouth, in 60 days; another from Port Maitland in 64 days; and one from Cockerwit Passage in 65 days (p. 875). However, it seems that the Cape Elizabeth groups swung east before reaching the Cape Ann line, because so many more of the former reached Nova Scotia than of the latter; i. e., that on the whole the two groups of bottles followed different routes until they converged toward the eastern side of the gulf.

The repetition, from year to year, of drifts most easily reconcilable with an anticlockwise eddying set argues strongly in favor of the prevalence of this type of circulation around the southern side of the basin of the gulf. Only one drift (No. 1773) from the two series so far launched off Cape Elizabeth (series A and E) has been hard to reconcile with this; because, if the date of recovery is correctly stated, its time interval from the offing of Cape Elizabeth to Grand Manan (56 days) is smaller than for any other bottle that crossed from the western side of the gulf to the Bay of Fundy. Granting it a direct journey, this means a daily rate of 2.7 miles, or at east 4.7 miles if it followed the eddying route, which is more likely.

The time intervals between the dates of release and recovery for bottles drifting from the offing of Cape Elizabeth to Nova Scotia averaged considerably shorter in 1923 (56 to 111 days; average 75 days for line E) than in 1922 (75 to 146 days; average 103 days for line A). Taken at its face value, this difference would point either to a more rapid rate of travel or to a more direct route, which in this case would mean veering more directly eastward. It seems more likely, however, that the difference is not as significant as it might appear, but that the discovery of the bottles and the local interest aroused thereby stimulated a closer scanning of the Nova Scotian shores in 1923, so that the bottles were found soon after they stranded, instead of lying on the beach perhaps for a week or more. The fact that one bottle, which drifted right up the Bay of Fundy to Advocate Harbor at Cobequid Point, at its head, was picked up in 107 days affords direct evidence to this effect, the distance on the assumed track being more than 250 miles.

With this uncertainty introducing a source of error that may be very considerable, I have not thought it justifiable to assume a shorter route for the bottles drifting to the mouth of the Bay of Fundy in 1923. The probable routes within the Bay of Fundy of such bottles from line E as entered the latter are laid down on the chart (fig. 182) to accord with the drift bottles set out there by Mavor in 1927 (i. e., crossing it from south to north and then continuing to veer westward to Grand Manan), because this type of circulation seems sufficiently established there.

Line E reproduces the corresponding series of the preceding year (line A), not only in the preponderance of drifts to Nova Scotia and in the uniformity of the tracks probably followed, but also in the recovery of one bottle at Metinic Island, off the western entrance to Penobscot Bay (No 1792), and of another at Round Pond Harbor, a few miles farther to the west (No. 1740). The time intervals for these (respectively, 64 and 77 days) correspond as closely as could be expected with 63 and 103 days for the two bottles (Nos. 98 and 284) that drifted to this same sector the year before (figs. 180 and 181), and hence suggest the equally circuitous offshore route laid down on the chart. However, it is possible that the two bottles in question (Nos. 1740 and 1792) actually circled in the opposite direction (i. e., clockwise), drifting inshore at first in company with four others that were picked up in Casco Bay and a few miles to the east of it, then continuing eastward along the coast, perhaps through the channels between the islands. The fact that one bottle (No. 1793) from the outer end of line E was found in Sheepscott River[76] after 34 days lends likelihood to this possibility.

The Cape Elizabeth series for the two years, however, illustrate an anual difference of another sort; namely, that the coastal belt, 10 to 15 miles broad next the cape, was a sort of deadwater in 1922 (p. 899), while in 1923 the general dominant set governed closer in to the coast.

BOTTLES SET OUT OFF MOUNT DESERT, AUGUST, 1923

The drifts of the bottles of the Mount Desert line can be approximated only if they are taken in conjunction with the several series discussed so far. Standing by themselves they would be self-contradictory, for 8 were recovered at significant distances to the westward (figs. 183 and 184); 11 were recovered at significant distances to the eastward; and 6 others at points close to where they were released. The easterly drifts so far reported all lead to the coast of Nova Scotia, except for one to the coast of Maine at the western entrance to the Grand Manan Channel (No. 1584, Haycock Harbor, Washington County). By themselves, these would naturally suggest a set to the northeast from the offing of Mount Desert, but analysis makes this most unlikely.

The fact that these Nova Scotian recoveries are distributed along the same sector of the coast line where bottles from the Cape Elizabeth, Cape Ann, and Cape

[76] Stated in the returns as "Sheepshead" River.

Cod lines have stranded would of itself be strong evidence that the routes of all had converged into one general and rather definite track some distance before they reached the land. In this respect the correspondence between the Mount Desert line of 1923 and the outer half of the Cape Elizabeth line of 1922 (series A, fig. 180)

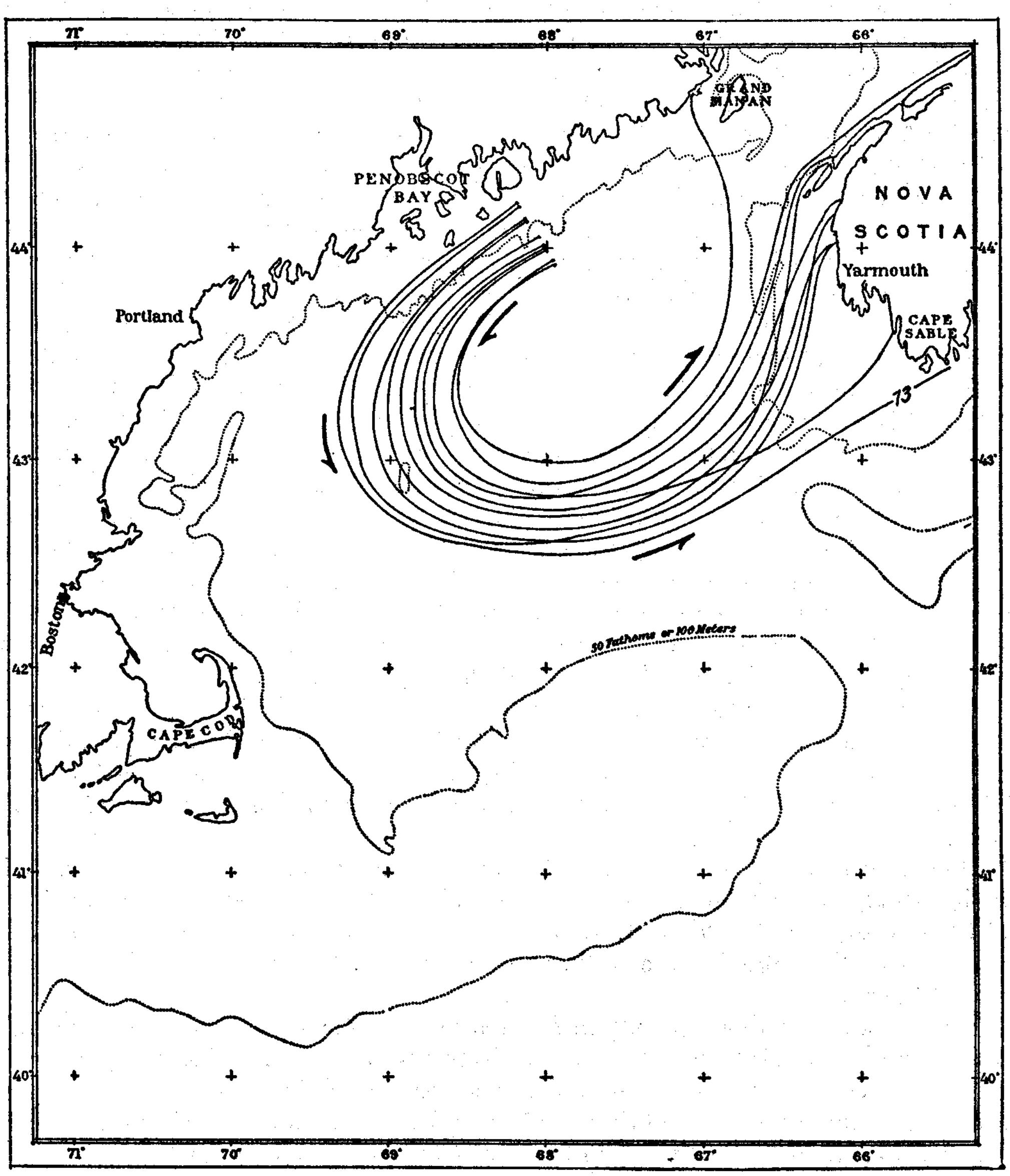

FIG. 183.—Assumed drifts of representative bottles recovered to the eastward from Series D, set out off Mount Desert Island, August 6, 1923. ●, places of release

is as close as could have been expected had all these bottles been launched on the same line and on the same day. In fact, this correspondence extends even to the odd bottles that diverged from the majority grouping, one or two having reached the vicinity of Cape Sable, and one or two found along the coast of Maine well to the eastward, in each instance. In the case of the drifts that cross the gulf, this track, I believe, is now definitely proven to approach the Bay of Fundy from the south or southwest, by the evidence just detailed.

The relationship which distance traveled bears to time interval between release and recovery also argues for a circuitous route for the bottles that went to Nova Scotia from the Mount Desert line, because the average distance for all of them, in a direct line, would be only about 85 miles, though the times range from 62 to 88 days for 8 of 10[77] (averaging 70 days). Evidence of this sort must, of course, be used with discrimination, because there is no knowing how long a bottle lies on the beach before it is noticed. When the results prove reasonably consistent, however, some trust can be put in them. In the present connection we have as a standard for comparison the Nova Scotian drifts from the lines set out off Cape Elizabeth. The distance (in a direct line) is only about one-half as great from Mount Desert to Nova Scotia as from Cape Elizabeth. The two lines of 1923 were set out only one day apart, and there is no reason to suppose that bottles from one line would be consistently overlooked while bottles from the other would be soon found. Consequently, it is reasonable to assume that some of the Mount Desert bottles would have been found a month or more before the first were reported from the Cape Elizabeth line, unless they had journeyed by a very circuitous route. Actually, however, the first four recoveries for the former were on October 7 to 9; the first three of the latter on the 9th and 10th. Allowing the one day's difference in the dates when the two series were put out, we have the rather surprising fact that the time intervals for these two groups, launched almost 100 miles apart, were the same almost to a day, though the strandings were scattered along more than 20 miles of coast line between Yarmouth, Nova Scotia, and the mouth of the Bay of Fundy.

The time intervals for the Nova Scotian drifts as a whole, from these two series, also correspond much more closely than the difference in direct distance would have suggested as probable, averaging about 75 days for the Cape Elizabeth series (extremes of 56 to 111 days; p. 875) and about 70 days for the Mount Desert group (62 to 151; p. 874).

The percentage of recoveries is not only of the same general order of magnitude for the Mount Desert line as for the Cape Elizabeth line of 1923 (respectively, 28 and 19 per cent), but the Nova Scotian and Fundian returns formed almost the same proportion of the total for the former (36 per cent of the total returns) as for the latter (42 per cent).

The most reasonable explanation for this correspondence between the two series, and the only explanation that fits all the facts just outlined, is that the journey to Nova Scotia covered about as long a distance for the Mount Desert bottles as for the Cape Elizabeth bottles, and that the former drifted southwestward at first, to join the general route of the latter group from west to east across the gulf.

Bottle No. 1584, set adrift about 25 miles out from Mount Desert Island and picked up at Haycocks Harbor, on the north shore of the Grand Manan Channel, 93 days later, probably followed the same general track as the bottles that went to

[77] Three others (Nos. 1530, 1551, and 1557), which were not picked up until 133 and 151 days had passed, may have lain unnoticed on the beach or drifted in and out along the shore with the tides.

Nova Scotia. It may have entered the south side of the Bay of Fundy, come out again past Grand Manan, and then circled the western end of the latter and so into the channel, as would be compatible with the current measurements in that region. Or it may have circled northward past the mouth of the bay but close enough to Grand Manan to be caught up in the indraft into the channel.

The general conclusion that all this group of bottles followed an eddylike course and did not drift directly eastward is directly corroborated by nine bottles from this same line, picked up to the westward along the coast of Maine. The fact that these were set out at intervals from the inner end of the line to the outer is evidence that the surface was involved in this movement for at least 25 miles out from the land.

Two bottles from the inner end of the line, picked up on Great Duck Island two days later, may have made their journey on the tide, for they were set out early in the ebb,[78] which sets toward the southwest here. A greater distance covered (10 miles) makes it likely that bottle No. 1515, which went to Long Island (also to the westward), made its landfall on the second tidal period; and it is certain that No. 1521, which went from the inner end of the line to Kennebunk, Me. (a distance of about 107 miles in a direct line), in 32 days, was carried with a very definite drift, for its rate was not less than 3½ miles per day. The daily rate of another bottle (No. 1523), which went from the mid section of the line to a point 8 miles southeast of Isle au Haut, 31 miles away, was ostensibly much more rapid, for it was reported as picked up the day after it was set out. This date, however, can hardly have been correct Allowing one day's error (which is probably the correct explanation), the daily rate would be about 7 miles to the westward.[79]

The rapidity of these westerly drifts, which can not be disputed, makes it likely that four other bottles that went from this line to the entrance to Penobscot Bay and to St. Georges River, a few miles farther west (Nos. 1553, 1565, 1566, and 1599), but were not found until after 35 to 38 days afloat, were drifting to and fro with the strong tides of Penobscot Bay for some days before they stranded and were noticed.

It is impossible, of course, to determine how far any given bottle, which moved westward from the Mount Desert line but did not soon strand, may have paralleled the coast before veering offshore toward the center of the gulf, but it is probable that most of them did so somewhere between the longitudes of Penobscot Bay and Cape Elizabeth. Had their general route led farther westward, more bottles from the Cape Elizabeth line might have been expected to show a southerly drift than the few actually so reported (p. 901).

Some few bottles from the Mount Desert line, hugging the shore line closest, may have crossed the Cape Elizabeth line, but the time intervals between release and recovery make it more likely that all that went across the gulf from the offing of Mount Desert passed to the seaward of the outer end of the Cape Elizabeth line—i. e., more than 25 miles offshore—and it is so indicated on the chart (fig. 182).

[78] It was high tide at Southwest Harbor at 6.26 a.m. on that day; the bottles in question (Nos. 1503 and 1510) were put out shortly afterwards.

[79] Assuming that it was picked up in the afternoon.

The tracks of three bottles from the mid section of line D, which were picked up at the eastern entrance to Frenchmans Bay, and one other that went to the vicinity of Petit Manan, are more puzzling. Ostensibly these point to short easterly drifts of 8 to 12 miles, and the time intervals are so uniform (33 to 38 days)[80] that all of them seem to have followed approximately the same route, though set out some miles apart. However, the time between release and recovery is so long for direct journeys so short, when contrasted with the rapidity with which other bottles set out near them drifted in the opposite direction, that it seems virtually certain that they followed a roundabout route. Judging from the facts that many more bottles stranded to the westward and that all of this series (D) were set out on the ebb, it is probable that the four bottles in question also drifted westward at first. Their most likely route would then be into Blue Hill Bay with the next flood, around Mount Desert Island, and so out again through Frenchmans Bay, to strand about Schoodic Promontory and to the eastward of it. Such a drift would be consistent with the clockwise circulation to be expected around Mount Desert Island, on theoretic grounds (p. 970). In short, the bottles set out off Mount Desert in 1923 afford definite proof of a set westward along the coast of Maine but no clear evidence of any longshore set in the opposite direction.

On the basis of the foregoing analysis, the most reasonable explanation of the localities where bottles from the Mount Desert, Cape Elizabeth, Cape Ann, and Cape Cod series of 1923 were recovered, and of the periods of time between the dates they were set afloat and later were picked up, is that bottles from all three lines moved in tracks eddying counterclockwise through southwest, through east, to north, and veering on successively shorter and shorter radii of curvature. Thus, the few bottles from the two southernmost lines, which were found on the Nova Scotian coast, probably traveled easterly from the time they were set out (southeast at first, then east and northeast), but the farther north and east along the coast bottles were put out, the more they tended to circle to the right of a direct course. It is also likely that while the breadth of the track covered by all the bottles in the western side of the gulf was something like 100 miles, they tended to converge into a narrower track as they approached the eastern side of the gulf.

In August, September, and October of 1922 and 1923 the center of this eddylike circulation seems to have been situated 40 to 60 miles south of Mount Desert Island, over the northeastern extension of the deep trough of the gulf.

The fact that the great majority of the recoveries from Nova Scotia and from the Bay of Fundy were from a rather short stretch of coast leads to the conclusion that no matter on which line the bottles in question were released, all those that drifted across the gulf finally came within the influence of the same south-north current, hugging close to the eastern shore. On no other assumption, I believe, is it possible to reconcile the facts just stated with the time element (p. 904) and with the current measurements that have been taken in that side of the gulf (p. 861).

The recoveries on the coast of Maine already discussed point to a division of this northerly set before it reaches the Bay of Fundy, the greater volume entering the bay along its southern shore, but offshoots (which may be only intermittent)

[80] No. 1511 was picked up in Winter Harbor 11 months later, a period so long that there is no way of estimating how far it may have traveled en route, or how long it may have lain on the strand.

from its western side recurving to the left across the mouth of the bay. Flotsam drifting in this branch may then come under the influence of the drift setting eastward into the right-hand side of the Grand Manan Channel. But only one bottle can so be classified, while five seem to have passed by the channel in their rounds to Penobscot Bay.

It is interesting that only two bottles from any of the several series[81] have been recovered along the coast sector between Petit Manan and the western entrance to the Grand Manan Channel, although many must have passed by. Judging from this, such parts of the dominant surface drift as veers westward past Grand Manan

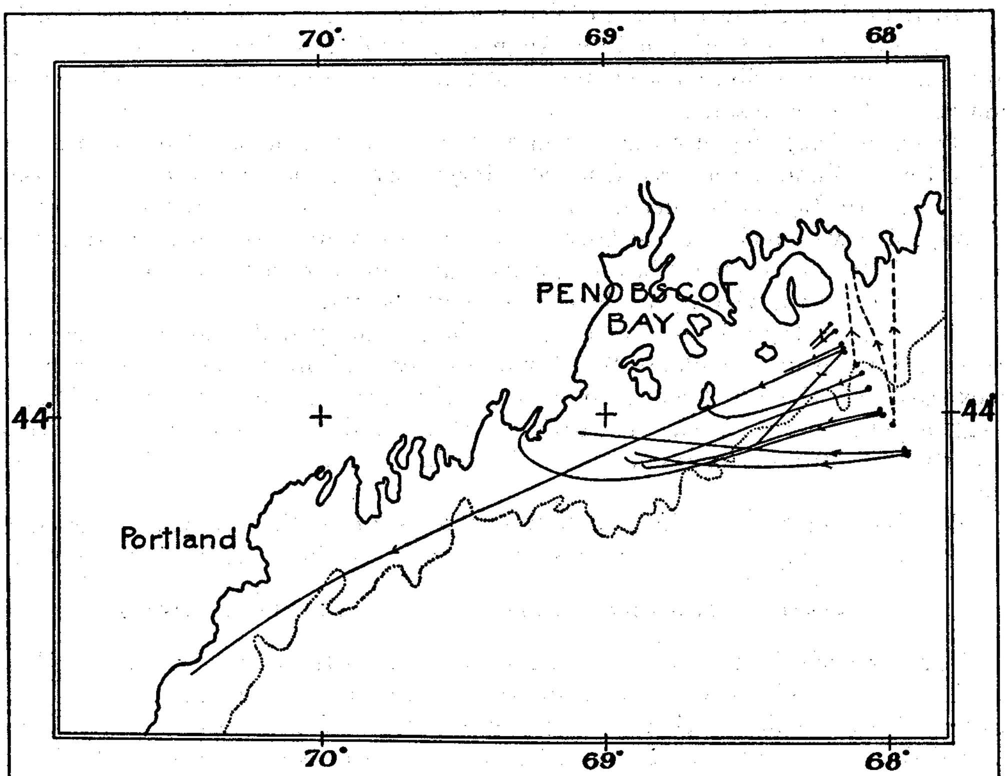

FIG. 184.—Assumed drifts of bottles recovered to the westward and inshore from Series D, set out off Mount Desert Island, August 6, 1923. ●, places of release

does not usually strike the coast of Maine in summer until it has passed the longitude of Mount Desert Island.

The circulatory scheme just outlined reconciles the bottle drifts for 1923 with those of the bottles set out in the Bay of Fundy in the summer of 1919, except that the latter certainly hugged the coast more closely in their westward and southward journey, else so many of them would hardly have been embayed behind Cape Cod. In this respect the summer of 1919 paralleled the April and July currents of 1925

[81] One set out in the Bay of Fundy in August, 1919 (Mavor, 1922, bottle no. 181); the other (no. 65) from the inner part of the Cape Elizabeth line of July, 1922.

and 1926, which carried bottles past Cape Ann, from Ipswich Bay into Massachusetts Bay (pp. 890, 893).

In the summers of 1922 and 1923 so many more bottles were picked up along Nova Scotia than in the western side of the gulf (a difference hardly accidental, because the coast line between Cape Elizabeth and Cape Cod is much frequented) that the surface water was evidently moving more offshore in the western side of the gulf, inshore in the eastern, than was the case in 1919.

BOTTLES PUT OUT OFF WESTERN NOVA SCOTIA

In 1926 the Biological Board of Canada put out four sets of bottles (each of 120) off Yarmouth, Nova Scotia, in July, August, September, and October, and Dr. A. G. Huntsman has contributed a summary of the recoveries in advance of his publication of the detailed results.

The great majority of returns from all the sets were from the Nova Scotian side of the Bay of Fundy, scattered from St. Marys Bay, at the mouth, to Minas Basin and Chignecto Bay, at the head. Six others crossed to the New Brunswick shore of the bay; five were picked up at Grand Manan; two went to the coast of Maine, one to Cape Cod; and two went in the opposite direction, eastward, past Cape Sable to Cape Negro and the vicinity of Shelburne, Nova Scotia.

As a whole, these drifts demonstrate the northerly drift along western Nova Scotia into the southern side of the Bay of Fundy, and up it. The New Brunswick recoveries show the anticlockwise movement within the bay, brought out by Mavor's (1923) experiments (p. 868). The drifts to Maine and Cape Cod are in line with the westerly and southerly drifts of bottles from the Mount Desert and Cape Elizabeth lines.

By what counterdrift the two bottles that went to the eastward escaped the Gulf of Maine eddy and came within the influence of the Scotian eddy is not clear.

DRIFTS OF BOTTLES ENTERING THE GULF FROM THE EASTWARD

The northerly drift along the Nova Scotian side of the gulf to the Bay of Fundy and its anticlockwise eddying continuation along the coast of Maine are further illustrated by the destinations reached by a considerable number of bottles that entered the gulf from lines set out off the outer coast of Nova Scotia by the Biological Board of Canada in the summers of 1922, 1923, and 1924. The following data have generously been contributed by Doctor Huntsman in advance of publication.

Two bottles from a line set out southeast across the continental shelf from Brazil Rock on July 17, 1922, were picked up along the western coast of Nova Scotia; 8 in the Bay of Fundy; and 2 circled farther westward, 1 of them to Winter Harbor and the other continuing past Mount Desert to the neighboring Long Island. The localities of release were scattered from 2 to 59 miles out from Brazil Rock, and none of the bottles set adrift farther out were reported from the gulf.

The bottle that went to Long Island made so rapid a drift (45 days from release to recovery) that no doubt it passed across the mouth of the Bay of Fundy. The Winter Harbor bottle, with 77 days, may have entered and circled the bay.

The next summer a series roughly at right angles to this last was set out on a line running northeastward from the western end of Browns Bank. Fourteen of these were reported from the Gulf of Maine; 6 of them scattered along the west shore of Nova Scotia; 7 were from widely separate localities in the Bay of Fundy, from its mouth to its head, after intervals of 64 days and upward; and 1 was from Penobscot Bay, Me. The drifts are thus much the same as those of the preceding summer, hugging the Nova Scotian coast to the Bay of Fundy. The time interval for the Penobscot Bay bottle was so long (113 days) that it, too, may have entered and circled the bay.

Twelve bottles from lines set out off Country Harbor, off Beaver Island, and off Cape Canso, Nova Scotia, were also reported from the Gulf of Maine (all of them from Yarmouth and northward along the western coast of Nova Scotia) and from the two sides of the Bay of Fundy. Only a single bottle from these eastern lines has yet been reported from the western side of the gulf—one set adrift a few miles off Sable Island by the Ice Patrol cutter *Tampa* on April 18, 1924, and picked up at Gloucester, Mass., 118 days later. The distance in a direct line being about 450 miles and something like 500 miles by the probable route around the northern side of the gulf, this bottle made an unusually rapid journey.

Since the preceding was written, Doctor Huntsman has contributed a summary of five monthly series, each of 200 bottles, set out offshore from Brazil Rock (off Cape Sable), July to October, 1926. Twenty-six of the 35 returns from the July set were close by, five from the Nova Scotian shore of the Bay of Fundy, and two from the New Brunswick side. The 46 returns from the August series were similar, except that the proportion of near-by returns was smaller (16); of returns from St. Marys Bay and the Nova Scotian shore of the Bay of Fundy larger (29). Fifteen of 23 returns from the lines of September 1 were again from this same sector of the Bay of Fundy region, three from the New Brunswick shore, and five between the point of release and Cape Sable. Three of the 15 returns from the series of September 27, however, were from the eastward (Shelburne to Port Mouton), nine near where set out, and only three from the Bay of Fundy. The series of October 20, again, gave 50 per cent of returns (six) near-by, four returns to the eastward (Negro Harbor to vicinity of La Have River), with two, only, from the western coast of Nova Scotia within the Gulf of Maine.

In sum, the evidence of a general northward drift hugging the Nova Scotian side of the gulf to the Bay of Fundy, and of its continuation westward as far as Penobscot Bay, is made cumulative by these drifts into the gulf.

The Brazil Rock series of 1926 also show the following seasonal succession: In July and August the dominant movement from the offing of Cape Sable was into the Gulf of Maine, but by the end of September the Scotian eddy had spread westward far enough to involve some bottles from this line in drifts best interpreted as circling anticlockwise, first offshore and then in again to the coast, to the eastward.

Further details as to the tracks followed are to be expected in Doctor Huntsman's forthcoming account.

One other bottle is recorded by the United States Hydrographic Office (Pilot Chart for May, 1923; reverse No. 26) as showing a similar drift into the eastern side of the Gulf of Maine from its release, 34 miles south of Cape Sable, September 21, 1902, to its recovery near Yarmouth, Nova Scotia, 30 days later.

CIRCULATION OF THE SUPERFICIAL STRATUM AS INDICATED BY SALINITY

The distribution of salinity affords a valuable check on the correctness of the circulatory system of the surface stratum, deducible from the drift-bottle experiments and from current measurements. The physical state of the water, together with the horizontal and vertical distribution of density, is the only clue yet available to the nontidal circulation in the deep strata of the gulf.

The reader will find frequent references to this phase of the subject in the section devoted to the salinity (p. 701). The distribution of salinity, as a reflection of the circulation of the gulf, has also been discussed in such detail in earlier reports on the Gulf of Maine explorations (Bigelow, 1914 to 1922) that a brief statement will suffice here.

With the oceanic water outside the edge of the continent much salter than the water in over the banks or alongshore (a rule prevailing all along eastern North America from Florida to the Grand Banks) a high salinity becomes an excellent indicator of any indraft from offshore. On the other hand, the lines of dispersal for land water are to be learned from the distribution of the least saline water. In the Gulf of Maine the flow of the Nova Scotian current past Cape Sable also tends to freshen the surface wherever its influence reaches.

Our first summer's cruise (in 1912) was enough to show what subsequent cruises have corroborated, that the freshest water is not localized off the mouths of the several large rivers, as would be the case if the discharges from these simply fanned out, but that it takes the form of a continuous and comparatively narrow belt skirting the coast line. The region where this freshest water does spread farthest out to sea (off Cape Ann and Massachusetts Bay) is some distance southward from the mouth of the Merrimac, the nearest of the large rivers. No fan of low salinity has ever been demonstrated off the mouth of the Kennebec.

The absence of such a fan off the mouth of any given river may or may not prove the failure of its discharge to drift out to sea, depending on the balance between the activity with which the tides mix the deep with the surface strata there and the volume of fresh water discharged. The river water that runs into the northern side of the gulf, and especially into the Bay of Fundy, is rapidly consumed in this way. Nevertheless, even where mixing is most active, areas of relatively lower salinity off the river mouths might be expected to alternate with areas relatively higher in salinity along the coast sectors between them, unless some dominant drift in one direction or the other disturbed this idealized picture. When we recall how great a volume of fresh water actually pours into the Gulf of Maine every year (p. 837) it is hardly conceivable that it would exert its chief freshening effect on so narrow a coastwise belt, unless the surface water tended to drift parallel to the land in the one direction or the other.

The summer salinities of 1912 (p. 770) pointed very clearly to a longshore movement of this sort around the northern and western margins of the gulf, setting westward along the coast of Maine, southward to Cape Ann, and spreading eastward off the cape in a rather definite tongue, outlined (at the surface) by the isohaline for 31.8 per mille (Bigelow, 1914, pl. 2). It was the presence of this tongue which established the direction of flow beyond dispute, because considerably higher salinities in Massachusetts Bay to the south of it, as well as offshore, left the coastal belt to the northward as its only possible source.

On the other hand, the salinity of the surface then afforded little evidence of river water in the northeastern corner of the gulf, in spite of the proximity of the St. John River. This, however, can be explained by the active mixing that takes place there, for while the mean salinity of the upper 50 to 60 meters was slightly higher (about 32.5 per mille) in the Grand Manan Channel and at its western end that August than it had been at the mouth of Massachusetts Bay a month earlier (about 32.2 per mille), the difference is no greater than can be explained as due to the regular seasonal succession (p. 799). A detailed discussion of the salinities for that summer, given in an earlier report (Bigelow, 1914, p. 90), leads to the conclusion that water of high salinity was being drawn into the eastern side of the gulf while the coastwise belt was dominated by a nontidal set alongshore from north and east to south and west, with expansions of water of low salinity off Penobscot Bay and off Cape Ann suggesting two separate anticlockwise eddies.

The subsequent summer cruises have expanded this preliminary concept of a general circling movement around the northern and western shores of the gulf to the domination of the surface over the entire basin by a great anticlockwise eddy, paralleling the land northward along Nova Scotia and swinging westward and then southward toward Cape Cod (Bigelow, 1917, p. 340), this being the only assumption on which the distribution of surface salinity can be rationalized.

This, it will be noted, has since been corroborated by the bottle drifts just described. A comparison between the recurving tongues of low salinity off Cape Ann and off Penobscot Bay, when such phenomena develop there, with the drifts from the Mount Desert, Cape Elizabeth, and Cape Ann lines, is especially instructive, for we find in such tongues a rational explanation for the tendency of the bottles to veer out from the land on successive radii. If, for example, bottles had been put out off Mount Desert in the summers of 1912 or of 1913, salinity suggests that the majority would have turned southward, abreast of Penobscot Bay, and that few, if any, would have stranded along the coast farther west. This actually happened in 1923 (p. 902, fig. 183). The tendency for bottles put out near land on the Cape Elizabeth and Cape Ann lines of that year to veer offshore from the beginning of their drifts would similarly find a reasonable cause in expansions of low salinity out toward the basin from the offing of Cape Ann, such as were actually recorded in July, 1912, and in August, 1914 (p. 763, fig. 136). But the distribution of surface salinity in August and September, 1915, when scattered observations outlined a band of low salinity of comparatively uniform breadth as paralleling the coast line from Nova Scotia to Cape Ann (fig. 137), would be compatible with drifts hugging the shore more closely as far as the cape, or perhaps to Massachusetts Bay, such as were actually followed by bottles set out in the Bay of Fundy during the summer of 1919 (p. 870) and off Cape Neddick (series O) in July, 1926. The locations of the isohalines at the surface are thus entirely reconcilable, both with the drifts assumed for the bottles and with the annual difference indicated by the sets put out in the summers of 1919, 1922, 1923, and 1926.

Mavor (1923), in his discussion of the distribution of salinities and temperatures in the Bay of Fundy for August, 1919, has shown that these are best explained as due to a movement of water into the bay on the Nova Scotian side, recognizable from the surface down to a depth of 100 meters, crossing northward toward New Brunswick about midway up the bay, with a counterbalancing outflow of water of low salinity southward and westward along the northern (New Brunswick) side. Here, again, temperature and salinity corroborate the evidence of drift bottles (p. 870).

The high surface salinities recorded in the northeastern corner of the gulf on the August cruises of 1912 and 1913 suggested a continuous tongue of highly saline water flowing into the eastern side of the gulf at the surface from the Atlantic Basin. However, subsequent discovery that the high surface values encountered in the basin between Maine and Nova Scotia in successive summers actually represent an isolated pool, resulting from local upwelling combined with tidal stirring (p. 768), and surrounded by less saline water on all sides, has lead to the appreciation that the gulf receives its saline water chiefly in the deeper strata (p. 842), not on the surface.

The rather abrupt west-east transition in surface salinity registered in the offing of Cape Sable in the summers of 1914 and 1915, added to the retreat of the critical isohalines (32 to 31.5 per mille) from the eastern side of the gulf, eastward, with the advance of the spring (p. 755), argues against any notable current from the east past the cape as characteristic of summer. Here, however, the effect which the active tidal mixing southwest and west of the cape would have in increasing the salinity of the surface, moving westward, must be taken into account.

If the evidence of salinity does not make clear the dominant set, if any, past Cape Sable for the summer months, the tongue of low salinity and low temperature found extending along the southeastern face of Georges Bank from northeast to southwest in July, 1914 (p. 770), is "hard to explain, except as an outflowing current from the gulf" (Bigelow, 1917, p. 241); and though this may not be a regular feature of the summer circulation (p. 608), the fact that several bottles from the Cape Cod and Cape Ann series of 1922 and 1923 seem to have drifted out of the gulf via this same route across the eastern end of Georges Bank (figs. 174 and 176) is certainly suggestive of its permanency. A tendency for water of low salinity to spread from the vicinity of Cape Cod southeastward to the neighboring part of Georges Bank is also indicated by the contrast in salinity between the western and eastern ends of the latter on the summer chart for 1914 (fig. 136, isohaline for 32.2 per mille). Here, again, a close parallel appears from the set, as indicated by the salinity of the surface water and the probable drift tracks of bottles that went in that direction from the Cape Cod series of 1922 (series B, p. 880, fig. 174). Farther south, in the southwestern part of the area, successive isohalines for 32.5 to 33.5 or 34 per mille, closely crowded and roughly paralleling the edge of the continent, prove that the dominant set here is along the outer part of the shelf, not transverse to it, though with some tendency indicated toward an eddying movement northward toward the land to the west of Nantucket Shoals. All this, again, is at once reconcilable with the drifts of bottles set out in this side of the gulf, especially with the tracks eddying westward out of the gulf past Nantucket Shoals, and with the group that went west from the edge of the continent abreast of Cape Cod (series B, outer end, p. 882).

The failure of any evidence, by salinity, of a surface drift from the continental edge out into the ocean basin in the region, in any summer of record, is corroborated by the fact that from the outer end of line B (fig. 174) only four bottles are known to have reached the general North Atlantic drift, and so to have gone across, one to England, one to Ireland, the other to the Canary Islands and the Azores.

The distribution of salinity at a depth of 40 meters has proved extremely diagnostic of the dominant circulation of the gulf, even more so than at the surface, the chart for July and August, 1914 (fig. 145), being the most instructive because covering the area as a whole. Its most noticeable feature—a continuous tongue of water of high salinity (33 to 33.4 per mille), extending from the Eastern Channel and Browns Bank inward to the north along the eastern side of the basin as far as the mouth of the Bay of Fundy—obviously reflects an unmistakable set of water into the gulf from the edge of the continent. The surface charts, the reader will recall, show nothing of this sort, evidence that the inward current (the existence of which is proven by several lines of evidence) did not involve the superficial stratum. Neither does it draw direct from the oceanic water (which would swing the isohalines for 34 to 35 per mille into the Eastern Channel), but from the mixture that takes place between tropic water and the water of the banks along the edge of the continent abreast of the gulf (p. 842). So far as the contour of the bottom is concerned, the whole southern aspect of the gulf, from Nantucket Shoals to the vicinity of Cape Sable, is open to overflows from this same source down to a depth of 40 meters.[82] Actually, however, we have found no evidence, in salinity, of any indraft of this sort anywhere to the westward of the Eastern Channel.

The expansion of the isohalines for 33 and 32.9 per mille to the westward along the coast of Maine, and the course of the isohaline for 32.5 per mille on the 40-meter chart just mentioned (fig. 145), combined with the location of the saltest tongue close against the eastern slope of the basin, are most readily reconcilable with a dominant set northward in the eastern side of the gulf (complicated by the evidences of upwelling in the offing of the Bay of Fundy already mentioned on p. 768), veering westward along the coast of Maine, and so southward around the periphery of the gulf, finally to turn southeastward as it is directed toward Georges Bank by the slopes of Nantucket Shoals.

This essentially reproduces the anticlockwise eddy indicated by the distribution of salinity at the surface (p. 911) as well as by the bottle drifts (p. 906), but the fact that the highest salinities at 40 meters lie 10 to 20 miles out from the 40-meter contour line in the eastern side of the gulf, not close in against the latter, is evidence that the eastern side of the eddy lay farther and farther out from the Nova Scotian

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[82] Except for the shoals on Georges Bank.

coast at increasing depths in 1914, as was also the case in August, 1913 (fig. 146). The comparative uniformity of salinity recorded over a wide area in the western side of the gulf at the 40-meter level in August, 1914, contrasted with the definitely outlined tongue of high salinity in the eastern side, points to the north-flowing side of the eddy as much more definite than the south-flowing side. In August, 1913, however, the distribution of salinity at 40 meters pointed to a closer approach to equality between the two sides of the eddy. The drift is not as clearly shown by the 40-meter salinities taken in August and September, 1915 (p. 990), except that the differential between higher salinities in the eastern side and lower ones in the western side of the gulf calls for some movement of the same anticlockwise sort, not being wholly explicable on the basis of upwelling, though assisted by that process (p. 768).

In none of these years (1913, 1914, and 1915) did the 40-meter level show the expansion of water of low salinity off Cape Ann that involved the upper 40 meters in July, 1912 (Bigelow, 1914, pl. 2; isohaline for 32.6 per mille at 25 fathoms), in a definite easterly drift. Thus, the distribution of salinity reflects much more variation, from summer to summer, at the 40-meter level in the western side of the gulf than in the eastern side, as well as at the surface (p. 770).

Unfortunately, the 40-meter chart for 1914 (fig. 145) does not so clearly show the dominant movement of water in the southwestern part of the area. However, isohalines closely crowded outside the 100-meter contour and the fact that they run parallel to the latter make it certain that no general drift was taking place transverse to the edge of the continent at the time, but that any dominant set that was then active roughly paralleled the latter. Consequently, the broad zone of 33 to 34 per mille between it and Nantucket Shoals (much more saline than any part of the Gulf of Maine at this level, but less so than the tropic water outside the continental edge) did not reflect a direct encroachment of the latter at the time or even any such movement earlier in the season, but merely reflected (by its precise salinity) the proportionate amounts in which water of higher and lower values had mingled there. However this may be, the presence of water of this comparatively high salinity to the south and southwest of Nantucket Shoals, added to rather an abrupt transition to considerably lower values (about 32.8 per mille) on the neighboring parts of Georges Bank, is good evidence that the surface drift, which has carried so many bottles out of the gulf westward across or around the shoals (p. 881), was not then operative to as great a depth as 40 meters, but that it is deflected more to the eastward, as the depth increases, by the contour of the bottom. This suggestion is corroborated to some extent by the fact that the isohalines for 33 per mille or lower include the whole eastern end of Georges Bank on the 40-meter chart in question, with an abrupt transition to much higher salinities (34.5 to 35 per mille) off its southeastern slope.

At first sight the presence of a tongue of water warmer than 10° running obliquely across Georges Bank from southwest to northeast at the 40-meter level, with lower temperatures within the gulf to the north as well as along the southeastern face of the bank (fig. 53), might seem to contradict this, but in this case salinity is the more reliable index to circulation, because the high 40-meter temperature at the station in question (10224), associated as it was with correspondingly low temperature (11.1°) at the surface, simply reflected active vertical mixing by tidal currents. Any tendency for the water to move from west to east over Georges Bank would necessarily be diverted by the considerable area shallower than 40 meters in which the bank culminates.[83] According to the rule general in the Northern Hemisphere, this shoal might be expected to act as the vortex for a clockwise circulatory movement, and the fact that the 40-meter salinity was somewhat lower on the eastern side of the bank than on the western side at the time, with the transition from values lower than 32 per mille to higher than 34.5 per mille most abrupt off its southeastern slope, is evidence of such a drift eddying eastward and southward around the shoal area.

The dominant circulation of the gulf is most clearly reflected in salinity at the time of year (spring and summer) when the regional variations in this respect are widest.

The progressive equalization of salinity that takes place during the autumn (p. 799) makes it increasingly difficult to reconstruct the horizontal circulation, even in its broadest aspects. In the midwinter of 1920–21 salinity yielded no definite evidence of any indraft into the eastern side of the gulf, either at the surface or at 40 meters (p. 804). It is unfortunate that observations could not be taken off Cape Sable during this midwinter cruise, for without such it is impossible to state whether the low values (31.2 to 31.3 per mille) recorded near Yarmouth, Nova Scotia, on January 4 (station 10501) reflected any movement of water past the cape from the eastward or were simply the product of local drainage from the land. However, it is certain that still lower salinity at the surface a few miles south of the Merrimac River, across the gulf, a few days earlier (30.02 per mille at station 10492) had the latter origin, and the rather abrupt transition appearing in both sides of the gulf on the surface chart (fig. 163) between water of low salinity (<31.5 per mille) close in to the land and considerably higher values (32.5 per mille) a few miles out at sea is definite proof that this coast belt was (or had been) drifting parallel to the shore line (if at all), not spreading inshore or offshore in either side of the gulf. However, the fact that the surface belt less saline than 32.3 per mille was much broader abreast of Penobscot Bay than in the offing of Casco Bay, on the one side, or off Mount Desert, on the other, points to some slight tendency for the water to drift out from the coast off the former, such as appears more definitely in the summer isohalines for 1912 (p. 770). Some such eddying movement is also indicated by the undulatory course of the isohalines off the mouth of the Bay of Fundy, suggesting a movement of water of low salinity out of its northern side toward the southwest, but no observations were taken close enough to the Nova Scotian side of the bay to develop the inward drift to be expected there.

The data for deeper levels were not distributed generally enough over the gulf during the midwinter cruise for safe interpretation in terms of dominant drift.

In early spring, when the discharges from the rivers increase, the courses of the isohalines become much more instructive with respect to the dominant drift, because they give a trustworthy clue to the lines of dispersal of the fresh water from the land. One of the most interesting phenomena in the hydrographic cycle of the gulf is the

[83] Minimum depth about 6 meters.

tendency of this water to hug the coast, not to fan out over the basin. At certain points along the coast local fishermen have long been aware that this results in a considerable southwesterly drift parallel with the coast, so much so that it is locally named the "spring current." The progressive development of a coastwise band of low salinity, which results from this event, is well illustrated by the successive surface charts for March and April, 1920 (figs. 91 and 101). Such distribution as appears on the latter and on the corresponding chart for May (fig. 120) could persist only with the water of the coastwise belt setting parallel to the general trend of the northern and western coast lines of the gulf, as already explained (p. 910). In the same way the expansion of water less saline than 32 per mille southward from the northern margin of the gulf, along its western shore, in a narrow band past Cape Ann to Massachusetts Bay, from March to April, and so out past Cape Cod toward Georges Bank by May (fig. 120), is unmistakable evidence of a general set of the surface water around the coast line along this same route.

The evidence afforded by salinity is therefore clear to the effect that when the outpouring of land water is at its maximum in spring it parallels the land, with a dominant flow alongshore from east and northeast to southwest and south, instead of spreading seaward, as happens off river mouths in many parts of the world. In other words, when the velocity of the left-hand side of the Gulf of Maine eddy is greatest it hugs the shore closest. The abrupt transition from surface salinity lower than 30 per mille to higher than 32 per mille, recorded 15 to 20 miles out from the western sector of the coast line between Cape Elizabeth and Cape Ann in May, 1915, gives a rough indication of the breadth of the zone along which the combined discharges from the Kennebec, Saco, and Merrimac Rivers are carried when the latter are in flood; and some indication that the main axis of this "spring current" is directed southward across the mouth of Massachusetts Bay, with some tendency to veer westward around the coast line of the latter after it passes Cape Ann, is traceable on the surface charts for April and May, 1925 (figs. 102 and 119), as noted above (p. 743). In this respect salinities and the drifts of bottles set out in Ipswich Bay (p. 890) prove mutually corroborative.

The charts of surface salinity for late summer for the several years, combined with the bottle drifts, suggest that the northern and western sides of the dominant eddy may be expected to trend more out from the land as the summer advances; but the isohalines point to considerable differences in this respect in different years, as just described (p. 770).

The chief line of dispersal for the discharge from the St. John River is located as tending toward the southwest past the eastern side of Grand Manan, by the sudden freshening of the surface recorded by Mavor (1923) at *Prince* station 3 from April to May in 1917 (p. 808; fig. 165), agreeing, again, with the routes probably followed by the bottles that drifted out of the bay in 1919 (p. 870); but the increase in salinity that takes place at this location from May to June and July is evidence equally positive that the velocity of this drift is at its maximum for only a few weeks (perhaps only a few days), though some movement of the surface water probably takes place in this direction throughout the year (p. 973).

The most interesting aspect of the seasonal dislocations of the isohalines in the southeastern part of the Gulf of Maine area is the light they throw on the fluctuations and lines of dispersal of the Nova Scotian current while this is flowing into the gulf from the eastward. The source of this cold water of low salinity and the chilling effect it exerts on the gulf are discussed in another chapter (p. 825), leaving for present consideration the rôle it plays in the dominant circulation of the gulf.

No dominant current of any great volume is demonstrable past Cape Sable in either direction from the salinities for August or September (though bottle drifts show the movements of water stated in another chapter—p. 908), nor in March (fig. 91); but when the Nova Scotian current commences to flood westward into the gulf in spring, its freshening effect is unmistakably reflected by a very noticeable dislocation of the critical isohalines (32 and 32.5 per mille). The seasonal schedule of this event varies from year to year, as described on page 832, 1920 being late in this respect, 1919 early, but experience in those years and in 1915 suggests that as the flow of the Nova Scotian current increases to its greatest head, it may be described as sweeping the isohalines westward before it far out into the gulf.

Unfortunately, our May cruise of 1915 did not extend to the southeastern part of the gulf, nor did the Canadian Fisheries Expedition take observations west of Halifax during that month, which leaves a wide gap for which I can not attempt to reconstruct the courses of the isohalines. However, the curves for 32 per mille salinity at the surface and at 40 meters (figs. 120 and 125) both outline the current as spreading westward from the cape toward the center of the gulf and somewhat fanlike toward the north.

This is corroborated by the fact that the *Grampus* encountered a strong set to the westward, upwards of 2 knots in velocity, on her run from the eastern side of the basin (station 10270) toward Seal Island, off Cape Sable (station 10271), on the 7th of that month. In that year, however, which may be taken as representative, the surface isohaline for 32 per mille had again withdrawn a considerable distance eastward toward Cape Sable by the last week in June (fig. 128), evidence that the westerly drift across the basin of the gulf ceased as soon as the flow of the Nova Scotian current slackened. The general distribution of salinity that characterizes the eastern side of the gulf in summer (p. 765) is best explained on the assumption that any water that rounds Cape Sable from the east during the months of July, August, and September veers northward along Nova Scotia toward the bay of Fundy, which is in accord with the drifts of the bottles that have entered the gulf from the east (p. 908).

It is obvious that the western extension of the Nova Scotian current must profoundly affect the nontidal circulation of water in the gulf at the season when it is at its maximum. Comparison of the surface salinity in May (fig. 120) with the currents deduced from bottle drifts in August suggests that this change consists chiefly in shifting the eastern side of the Gulf of Maine eddy westward—how far, can not yet be stated.

It is also obvious that if the the anticlockwise eddy persists through spring (as there is ample evidence, theoretic as well as direct) it must constantly draw into its eastern side (and so carry northward toward the Bay of Fundy and the coast of

Maine) an admixture of the colder and less saline water from the Nova Scotian current; but the details of this process and the extent to which it influences the temperature, salinity, and circulation of the northeastern part of the gulf can not be worked out until more data are gathered for the critical months of May and June.

It is much to be regretted that no records on the eastern part of Georges Bank have been obtained for June, which might throw light on the expansion of Nova Scotian water in that direction; but the fact that we have found the surface salinity considerably higher in the Eastern Channel and in the basin of the gulf near by than from Browns Bank in to Cape Sable, both in June and in July (figs. 128 and 136), shows that any movement that may take place along this zone toward the southwest in spring had ceased by the beginning of the summer both in 1914 and in 1915.

CIRCULATION OF THE SUPERFICIAL STRATUM AS INDICATED BY TEMPERATURE

The distribution of temperature is by no means as clear an index to the nontidal circulation of the surface waters of the gulf as is its salinity, because any given mass of surface water may be warmed rapidly by the sun or cooled by radiation when the overlying air is the colder without suffering any alteration in its identity by mixture with other water masses. In the deep strata, however, which are more or less insulated from these thermal influences from above, regional differences in temperature are more easily interpreted in terms of circulation.

The relationship of temperature to circulation is referred to repeatedly in other connections;[84] only the most salient aspects, then, need be referred to here.

The belt of coldest water, which fringes the shores of the gulf in winter, owes its low temperature to the chilling effects of the icy winds that blow out over it from the land. The fact that this cold band (as illustrated by the surface charts for midwinter (fig. 80) and for February to March—fig. 1) is comparatively uniform in breadth all along the northern and western shore line, is best reconciled with a set paralleling the shore. Any considerable movement of surface water either from the land out to sea or vice versa would give much more undulatory courses to the critical isotherms of 5° in December to January and 2° in February to March.

Surface water equally cold over the Northern Channel and Browns Bank on the February to March chart (fig. 1), giving place, by a rather abrupt transition, to readings 1.5° higher over the Eastern Channel, reflects the westernmost bound of the Nova Scotian current at the time; and an expansion of water colder than 4° out over the channel from the the gulf and across the eastern end of Georges Bank but not across the western end of the bank is evidence of a movement in that direction, which corresponds to the drifts of bottles set out off Cape Cod in summer (p. 886).

The undulatory course of the March isotherm for 3° gives a rather clear indication of an anticlockwise eddying movement in the central part of the gulf, with warmer water moving northward in the eastern arm of the basin and colder water drifting out from the land off Penobscot Bay, illustrating one of the varying forms

[84] See the chapter on temperature.

of the Gulf of Maine eddies. This same distribution of temperature, however, reappearing in April, is reminiscent of a past state of circulation, not of a present one, because the corresponding charts of salinity show the dominant set to have assumed a southwesterly course, more nearly parallel to the coast line, from the one month to the next (p. 743). Neither of these early spring charts of temperature suggest any drift of warmer water into the eastern side of the gulf from offshore; but some drift of this sort is indicated on the 40-meter chart for March (p. 525) by a band warmer than 3° entering via the Eastern Channel. This indraft appears more clearly at deeper levels (p. 526).

With the advance of spring the regional inequalities of temperature become increasingly significant, from the standpoint of circulation, as they outline the lines of dispersal followed in the gulf by the cold water of the Nova Scotian current. In general, temperature corroborates salinity to the effect that the current did not begin to flood westward past Cape Sable until after the middle of April in the year 1920, though it had exerted its chilling effect in this direction as far as the eastern side of the basin of the gulf by the last of March the year before (p. 553). The isotherms for May (fig. 27), however, suggest more of a tendency for this Nova Scotian water to spread northward toward Maine and the Bay of Fundy, as well as westward in the gulf, when at its head, than do the isohalines (p. 745).

Rising temperature, like rising salinity, reflected a slackening in the current in 1915 from May to the last half of June, when an abrupt transition in the temperature of the coldest stratum, from the Eastern Channel (about 8.1°) to the vicinity of Cape Sable (about 0.7°), located its southwestern boundary at Browns Bank. This is also indicated by the abrupt transition from colder to warmer water along the western slope of the bank at 40 meters; but the low temperatures recorded over the southwest slope of Georges Bank on the July profile for 1914 (fig. 58, p. 616)[85] is readiest explained as reminiscent of a cool current skirting the bank from northeast to south some time previous. It seems that in the cold year 1916 such a drift of cool water was either in much greater volume or persisted until later in the season, for it is difficult to account otherwise for the band of low temperature which the *Grampus* encountered over the southwestern slope of the bank that July (p. 629).

"The facts that the cold band of 1916 lay almost exactly in the prolongation of that of 1914; that a similar streak of comparatively low temperature (6.4°) was encountered at the same relative position on the shelf some 60 miles farther west in 1913 (station 10062); and that the axis of the coldest water noted on the shelf south of Nantucket in 1889 (Libbey, 1891) merely prolongs this general zone, practically amount to proof that a northeast to southwest flow of cold water takes place there annually in late spring or summer, dovetailing in between the warmer and fresher bank water on the north and the Gulf Stream on the south." (Bigelow, 1922, p. 166.) Its source is discussed elsewhere (p. 848). The July isotherms for 1914 locate its extreme western boundary between longitude 68° and 69°, where the 40-meter chart

[85] This also appears on the corresponding chart for the 40-meter level, but is complicated there by active vertical mixing that maintains a higher temperature over the shoal parts of the bank at this depth (lower at the surface) than on its southern side; the alternation of a warm with a cold belt along the bank, outlined in the 40-meter chart (fig. 53), is therefore partly of local origin.

(fig. 53; isotherms for 10° and 12°) suggests an eddying movement, drawing warmer water inward over the bank on the western side; but in other summers the cool drift extends much farther westward. Bottle drifts, for example, place 1922 in this category (p. 883); and Libbey (1891 and 1895) records it in longitude 70° to 71° in the summer and early autumn of 1889.

In another chapter (p. 585) I have tried to make it clear that the areas of low and high surface temperature, which characterize various parts of the Gulf of Maine in summer, are due chiefly to tidal stirring—most active over the shoal banks and in the northeastern part of the area generally, least so in the basin off Massachussetts Bay. Tidal stirring also plays a part in holding the surface temperature somewhat lower along the western margin of the gulf and around the shore of Massachussetts Bay than a few miles out at sea; but the gradation also points to some movement of the surface water eastward, away from the shore, under the impulse of the prevailing southwestern winds, an event with which bathers on our beaches have long been familiar (p. 588), and which takes part in the development of the western side of the Gulf of Maine eddy. The evidence (by bottle drifts) of a westerly set from the Nova Scotian side and from the Bay of Fundy along the coast of Maine is also borne out by the extension of surface water colder than 14° westward past Penobscot Bay in August (figs. 46 and 47) over depths so great that tidal stirring, *in situ*, is not active enough to be responsible, *per se*, for surface values as low as those actually recorded there.

The 40-meter charts for July and August (figs. 52, 53, and 54) also suggest a similar westerly drift by the isotherms for 8° and 9°, though at this depth the water moving in that direction from the Nova Scotian side is warmer than that which it replaces off the coast of Maine—not colder, as it is at the surface. or discussion of this bathymetric difference, see p. 608).

The mutual relationships of waters warmer and colder than 9° were especially suggestive in August, 1913, as locating the vortex of the anticlockwise eddy about 60 miles south of Mount Desert and Penobscot Bay (fig. 52). The corresponding chart for 1914 (fig. 53) is not so easy to interpret in this respect, the picture being complicated in the western side by a pool of water cooler than 6°, which probably owed its low temperature to vertical stirring or to local upwelling in the mid depths.

None of the summer charts for temperature reveals any dominant movement of warm water into the gulf from offshore at the surface, nor do the 40-meter charts for the summers of 1914 or 1915, but some circulatory indraft of this sort is suggested on the 40-meter chart for 1913 (fig. 52) by temperature, just as it is by salinity (p. 782), by the warm (>10°) tongue in the eastern side of the basin, with lower temperatures on either hand, to which the reader's attention has already been called (p. 608).

At first sight the distribution of temperatures at 40 meters prevailing in July, 1914 (fig. 53), might suggest a drift into the gulf from offshore across the eastern end of Georges Bank, but a closer analysis makes it clear (p. 617) that in this case unity of temperature had a local significance only, being an adventitious result of the fact that vertical mixing was most active on the northern part of the bank.

CIRCULATION IN THE DEEP STRATA AS INDICATED BY TEMPERATURE AND SALINITY

Dawson's (1905) observations made it known that the tidal currents of the eastern side of the gulf run about as strongly down to a depth of 55 meters as they do at the surface, and measurements taken at 5 stations by the *Grampus* in the summer of 1912 showed bottom currents varying in velocity from 0.1 to 0.25 knot per hour in depths of 100 to 265 meters (Bigelow, 1914, p. 86). Evidently, then, the basin of the gulf is constantly in a state of active circulation right down to the bottom, its whole mass of water oscillating to and fro with the tides, though with velocities somewhat lower in the deep water than at the surface.

Up to the present time no attempt has been made to determine the nontidal movement of the bottom water of the gulf with current meters or by the use of deep drift bottles, such as have proved so instructive in the North Sea, but the regional differences in temperature and salinity outline the major movements over the bottom.

At depths greater than 100 meters the gulf of Maine is an inclosed basin with the narrow Eastern and Northern Channels as the only possible entrances or exits through which water can flow in or out of its basin. It follows from this that any deep current into the gulf can enter only in its eastern side. Such entrance might be via either of the two channels or through both, so far as the contour of the bottom is concerned. Actually, however, salinity and temperature show that the indraft of slope water over the bottom is restricted to the Eastern Channel, the abrupt west-east transition in salinity and in temperature, which characterizes the Northern Channel, being incompatible with any large transference of bottom water through the latter in either direction.

The dominant drift in the eastern side of the Eastern Channel is clearly northerly (into the gulf) at all times of year, but a considerable difference between high values of temperature and salinity in the eastern side of the channel and lower values in its western side in March, April, and July (pp. 770, 789) point to an outflowing current via the latter, continuing southward and westward around the slope of Georges Bank.

Slope water is betrayed in the deep strata of the gulf by its high salinity (33.5–34 per mille, p. 849) and moderately high temperature (4.5° to 8°). At the 100-meter level the isotherms and isohalines show the inflowing current hugging the eastern slope of the basin in March as a rather definite tongue of high temperature and salinity (figs. 13 and 94), veering westward around the northern side of the basin, with a countermovement of cooler and less saline water setting southward and eastward around the southern side of the basin. In fact, physical evidence could hardly be clearer that the general Gulf of Maine eddy was effective to a depth of at least 100 meters in this particular month, though complicated by an indraft through the Eastern Channel in the deeper levels, which did not directly affect the surface (p. 704).

An anticlockwise circulation is also indicated on the 100-meter charts for April (figs. 25 and 116), though less clearly, by concentration of the highest salinities and temperatures in the eastern and northern parts of the basin, the lowest in the western and southern parts. In this case, however, the westerly component involved a broader and less definite band off the coast of Maine than in March, and the easterly component of the eddy had shifted southward to skirt the northern slopes of Georges Bank more closely.

Information as to the movement of water along the bottom of the Northern Channel is much to be desired at the season when the Nova Scotian current is flooding in greatest volume into the gulf. Some drift may be assumed to take place into the gulf by this route as deep as 100 meters in 1915, to account for the concentration of the most saline water in the western side of the basin at the 100-meter level in May (fig. 127), instead of in the eastern side, as at other times of year. It is probable, therefore, that when the drift past Cape Sable is at its maximum it causes a westerly shift in the vortex of the general eddy in the mid depths, though not essentially altering the anticlockwise type of circulation, however. Any westerly drift that may have taken place along the bottom of the Northern Channel in 1915 had ceased by June; on this basis, alone, is the abrupt east-west transition that appears there on the 100-meter chart of temperature for that month explicable (fig. 43).

In midsummer the transition from lower salinities and temperatures in the western side of the gulf to higher in the eastern, at the 100-meter level, and the sweep of the successive isohalines and isotherms from east to west along the northern slope of the gulf, again give evidence of a general set northerly past Nova Scotia and westerly along the coast of Maine in the mid depths, paralleling the dominant circulation at the surface. The nontidal movement of water of the southern side of the basin at this level is not so clear, the picture being confused by an area of relatively high salinity and temperature off the northern slope of Georges Bank near the entrance to the Eastern Channel, which is not easy to account for.

In spite of this and of other apparent anomalies the distribution of temperature and salinity in the mid depths, as exemplified by the 100-meter level, are, as a whole, compatible with the domination of the basin by the general Gulf of Maine eddy, anticlockwise in character.

The horizontal circulation of the gulf at greater and greater depths is more and more directed by the contour of the bottom, which gives the basin the outlines of a Y, with two arms uniting and open to the Eastern Channel (p. 784) at 175 meters, but entirely inclosed at 200 meters and deeper.

With temperatures and salinities recorded at one deep station or another for so many months and years, it can be stated confidently that the movement of bottom water inward into the gulf takes place in pulses, the secular fluctuations of which have only been glimpsed as yet (p. 850). On the other hand, dynamics (fig. 204) and the distribution of temperature and salinity point to some outgoing drift via the western (Georges Bank) side of the Eastern Channel between these pulses in summer (pp. 789, 852).

The presence of water of high salinity (34 per mille) in both arms of the trough but never (so far as yet recorded) over the submarine ridge that separates them is good evidence that the latter divides the slope water as it drifts inward in the deepest stratum of the gulf.

Two separate anticlockwise eddying drifts are indicated in the bottoms of the two arms of the trough, at depths of 175 meters and deeper, by salinities and temperatures averaging somewhat higher on the side that would be to the right, for an inflowing current, than on that to the left (p. 785). The circulation in each may therefore be described as "estuarine," subsidiary to the estuarine circulation of the basin of the gulf as a whole, inward along the right-hand (eastern and northern) sides and eddying to the left. The regional difference between the right and left sides being widest in the eastern trough, with the maximum values of salinity and temperature both higher there than in the western, a greater volume of slope water continues northward over the bottom toward the Bay of Fundy (and at a greater velocity) than is diverted to the westward by the ridge that culminates in Cashes Ledge.

CIRCULATION AS INDICATED BY THE PLANKTON

The tracks which immigrant members of the planktonic community follow into the gulf and in their further wanderings within it are discussed in such detail in the preceding number of this volume (Bigelow, 1926, p. 51), to which the reader is referred for details, that the briefest of summaries will suffice here. Immigrants of this category, whether from tropic or from northern sources, enter the gulf in the eastern side; seldom or never across its offshore rim farther west. (Bigelow, 1926, figs. 31 32, 33, 69, 71, and 72.) The relative regional abundance of our northern copepods, *Calanus hyperboreus* and *Metridia longa* (Bigelow, 1926, figs. 71 and 76), clearly pictures the drift westward into the gulf from the offing of Cape Sable and westward along the offshore slope of Georges Bank in the spring; and the records for the more delicate northern visitors—Mertensia, Ptychogena, *Oikopleura vanhöffeni*, and *Limacina helicina*—are chiefly confined to the area on the eastern side, where the water is most chilled by the Nova Scotian current.

Clearest evidence of the drift within the gulf is afforded, of course, by such species as are comparatively short lived there and can not reproduce in its low (or high) temperature. The records for these in the upper 40 meters or so have been constantly confined to a rather definite belt paralleling the coast around from the Nova Scotian side to the offing of Massachusetts Bay, leaving the central and southern parts of the gulf bare (Bigelow, 1926, fig. 31). A distribution of this sort is reconcilable with an eddying drift inward, anticlockwise around the gulf; in fact, it is explicable on no other reasonable assumption, and this corroborates the drift-bottle experiments. A drift of this same sort from the coast of Maine westward and southward toward Cape Cod is also made probable by the relative distribution of buoyant fish eggs and of larval fishes (Bigelow, 1926, figs. 34 and 35). Planktonic animals that enter the gulf in the mid levels via the Eastern Channel (*Eukrohnia hamata*, for example) parallel the surface communities in their general drift northward, westward, and southwestward, except that they are held farther out in the basin by the contour of the bottom; but visitors characteristic of the deepest water of the gulf (e. g., *Sagitta maxima*) follow the two arms of the Y-shaped trough, just as might be expected from the drift of the slope water, as indicated by the salinity (p. 922).

The comparative scarcity of animals of coastwise or shoal-water origin over the deep basin of the gulf (Bigelow, 1926, p. 32), like the distribution of salinity, is evidence of a circulatory system paralleling the coast, not fanning out in the offing of the river mouths.

VERTICAL STABILITY AS AFFECTING THE CIRCULATION OF THE GULF

A clue to the relative strength of vertical currents in different parts of the gulf during the warm months is afforded by the relative degree of vertical stability of the water that opposes them.

The relationship between vertical circulation and stability is simple. Whenever or wherever the water is so nearly homogeneous as to the density that it has little or no vertical stability (as is the case in the coastwise belt of the Gulf of Maine in winter), vertical mixings or upwellings freely follow the tidal circulation and the disturbing effects which the wind exercises on the surface; but if the superficial stratum be made much lighter than the underlying stata by freshening or by solar warming, it requires a considerable expenditure of force to drive the light surface water down or to bring heavy water up from below. It is conceivable, also, that the column might become so stable as to effectually insulate the deeps from any influence from above.

The activity of vertical circulation at any time or place in the gulf, therefore, depends on the momentary balance between the mixing tendency of the tides, etc., and the degree of vertical stability by which this is opposed.

It is important to bear in mind that any given particle of water has no stability *per se*, but only relative to the water above and below it. It is usual, therefore, in hydrodynamic calculations, to state the stability for strata of convenient thickness.[86] Being strictly a function of the density of the water, a simple visual measure of its relative value is afforded by the usual curves for density, plotted against depth, remembering that the more the curves depart from the vertical, the higher the stability, and that it is zero throughout any stratum where the curve is vertical.

Regional variations in this respect may be represented graphically by plotting the differences in density between the surface and some underlying stratum chosen as a base, as in Figure 185. The greater the difference, the the more stable the water.

In the Gulf of Maine the tidal currents are strong enough at all depths to effect an active mixing of the water, were they unhindered; and the consumption of slope water that takes place in the inner part of the basin (p. 941), with its constant replenishment from offshore, is unmistakable evidence of some interchange between surface and bottom. The prevalence of a decided contrast in salinity between the superficial and deep strata throughout the year proves this interchange a slow process, however, wherever the water is more than 100 meters or so deep. The limiting factor here is the stability of the water, for the specific gravity of the slope water in the bottom of the gulf is always considerably higher than that of the superficial stratum, even in winter, when the latter is heaviest and itself has little or no stability.

The gulf as a whole, then, is always in a state of stable equilibrium, whatever may be the state of the water near its surface; and while not sufficiently so to prevent vertical mixing from taking place constantly, we have no record of slope water welling up to the surface from the bottom of the trough, nor is such an event to be expected.

[86] The unit of stability usually employed is the number of surfaces of equal specific volume per 10 meters of depth, represented graphically by vertical lines varying in breadth according to the stability of the water in the several strata. (Sandström, 1919, p. 283.)

The vertical stability varies little from season to season in the bottom stratum deeper than 100 meters, indicating comparative uniformity in the activity of vertical circulation there; but wide seasonal fluctuations in the stability of the superficial stratum reflect corresponding differences in the stirring effects of the tides, etc.

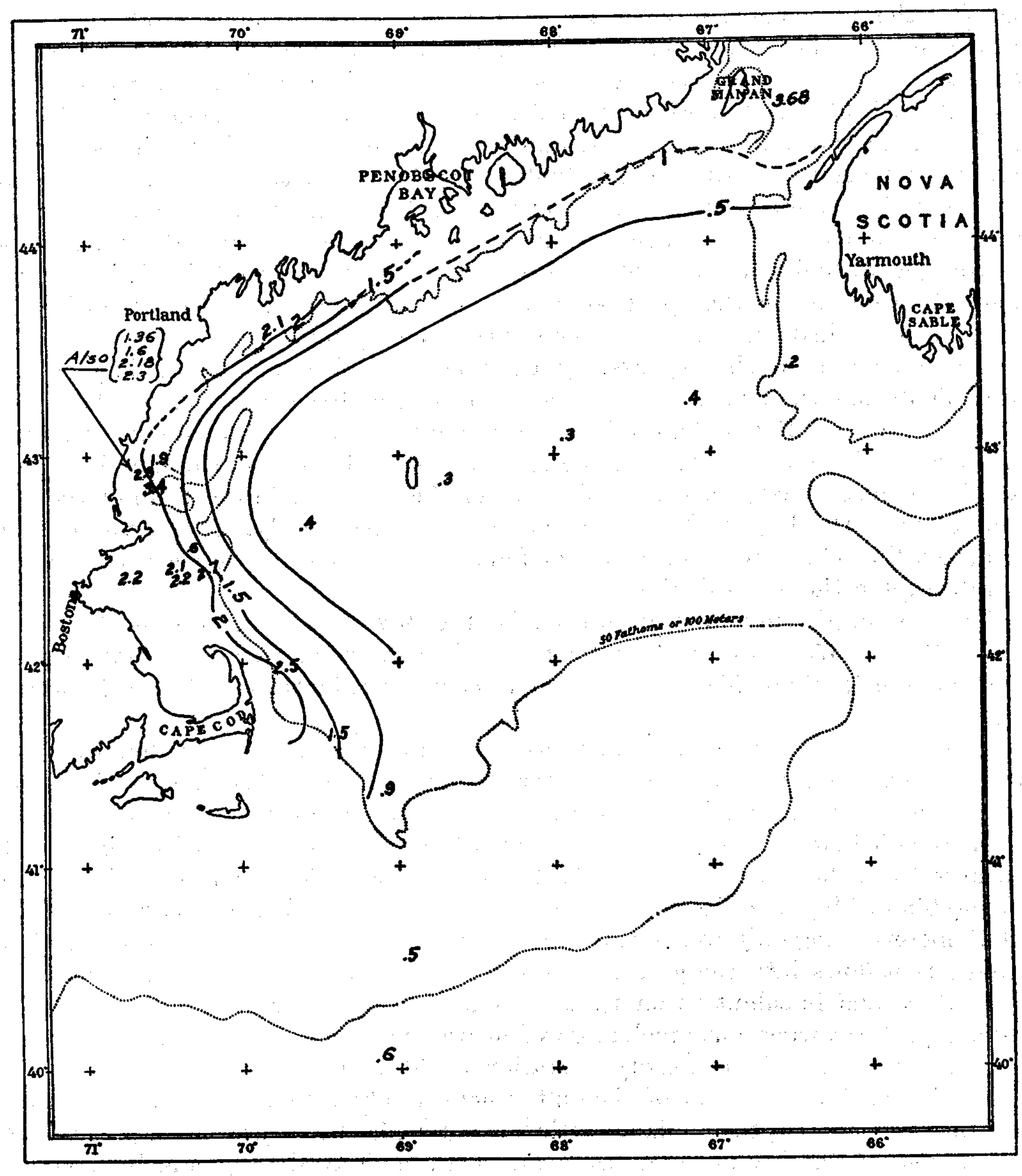

FIG. 185.—Difference in density between the surface and the 40-meter level for May, 1915 and 1920. Corrected for compression

In northern coastal waters generally (the Gulf of Maine is no exception) vertica mixing of the upper 100 meters is most active during the coldest season, when, thanks to low surface temperature, the water has little stability; and it is at this season that the consumption of slope water is most rapid. From March to April, however, vertical currents in the coastal belt between Cape Cod and Cape Ann are suddenly opposed by the increase in stability effected by the combined effect of the freshening of the surface by the river freshets and of the rising surface temperature. Together these processes produce an average difference of about 2 to 3.5 units of density between the surface and the 40-meter level in this zone by May (fig. 185). The most stable state yet recorded in the gulf was off the mouth of the Saco River on April 10, 1915 (Bigelow, 1914a, p. 417), when very low surface salinity (26.74 per mille) was responsible for a vertical range of 4.53 in density within this stratum, showing that vertical mixings had virtually ceased, for the time being. May also sees the rather sudden establishment of a high degree of stability in the Bay of Fundy consequent on the sudden lowering of the salinity of the surface by the freshets from the St. John River (p. 808), Mavor (1923) having recorded a difference of about 3.7 in density in the upper 40 meters on May 4, 1917, at *Prince* station 3, where the water had been virtually homogeneous on April 9.

The Penobscot freshet apparently has much less effect on the stability of the water off its mouth; and without sufficient inrush of fresh water along the coast between Penobscot Bay and Grand Manan to offset the active tidal mixing, we find that in May the upper stratum of the gulf is most stable in its two opposite sides, viz, Massachusetts Bay to Cape Elizabeth in the west and in the train of the St. John River in the Bay of Fundy in the east. Consequently, the active vertical circulation that characterizes the Bay of Fundy during most of the year is temporarily interrupted there at this time.

This period of temporary quiescence for the Bay of Fundy is of brief duration, Mavor (1923, p. 375) showing the 40-meter stability decreasing again by June to only about one-fourth of the May value as the river water is incorporated into the water of the bay.

I can not state the stability along western Nova Scotia for May; but it is not likely that the small amount of fresh water emptying in along this sector of the coast line can offset the active mixing which the strong tidal currents tend to effect there.

In the offshore parts of the gulf, to which the freshening effect of the increased discharge from the rivers has not yet extended, the superficial stratum is but little more stable in May than in April, the average difference in density between surface and 40 meters rising only to about 0.3 over the basin generally. The Nova Scotian current, as it flows into the gulf from the east, is so nearly homogeneous, both in temperature and in salinity, that it, too, is but slightly stable, though considerably lighter than the warmer but much more saline water in the eastern side of the trough over which it floats (cf. the density at station 10270, p. 988).

In the southwestern part of the gulf generally, where tidal currents are weaker than in the northeast, their mixing action is not sufficient to prevent a progressive development of stability in the upper 40 meters through April, May, and June as the surface warms; and as soon as the surface temperature has risen appreciably above that of the underlying water, upwellings are readily recognized by their chilling effect.

As remarked in another chapter (p. 550), water often wells up from below along the western side of the gulf in spring, when offshore gales drive the surface water out to sea. Bathers on New England beaches also are familiar with this same event in summer (p. 588). The fact that the surface averages somewhat cooler along the coast at that season, from Cape Cod to Cape Elizabeth, than a few miles offshore probably reflects the cumulative effect of such upwellings following the prevailing southwesterly

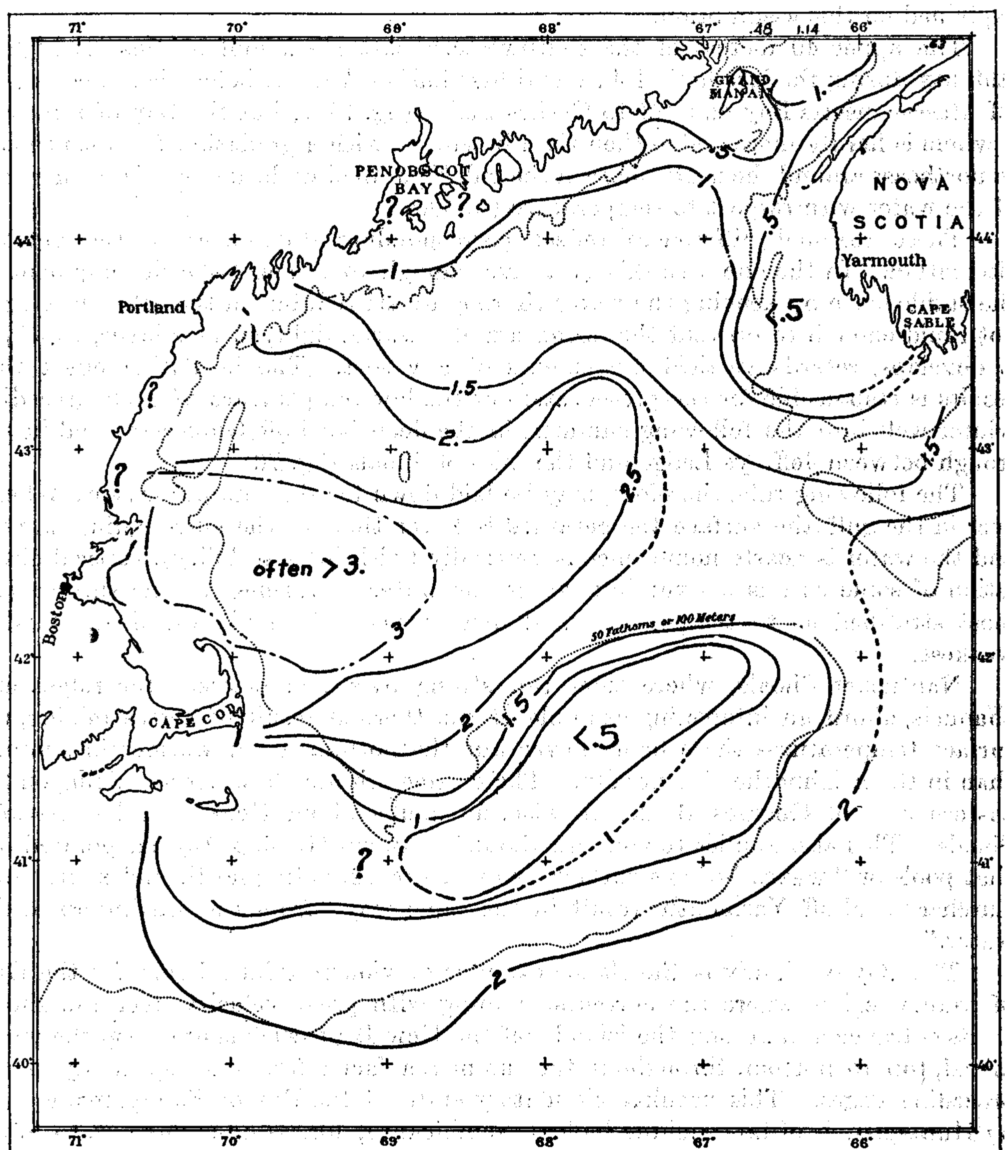

FIG. 186.—Difference in density between the surface and the 40-meter level in July and August for the several years of record, combined. Corrected for compression

winds. No doubt this happens still more frequently there in winter, when northwest gales are frequent, though it is not so easily recognizable then. In the opposite side of the gulf the tendency is the reverse—i. e., the surface water is driven in against the shore and sinks; and with vertical mixing by the tides so active that but little stability develops there, more or less overturning of this sort probably takes place along the coast of Nova Scotia even at the warmest season. The frequency with which bottles have stranded there after drifting across the gulf may be explained on this assumption.

The upper 40 meters of the southwestern part of the gulf attains its highest stability during the last half of July and first half of August, being then most stable off Massachusetts Bay and out to Cashes Ledge (fig. 186); but the Bay of Fundy as a whole is hardly more stable then than in winter, with a gradation from southwest to northeast around the north shore of the gulf,[87] paralleling the degree of stratification of the water with respect to temperature (p. 596).

These regional differences reflect corresponding differences in the vertica circulation. In the one case this is active enough to prevent the development of the stable state by keeping the water thoroughly stirred, but in the other mixing is not rapid enough to prevent the formation of a warm, light, surface layer, which, as it develops, retards vertical movements of any sort. The insulating effect that results is responsible for the preservation of the low temperature of each preceding winter well into the following summer, in the deep bowl off Gloucester and in the trough between Jeffreys Ledge and the Isles of Shoals (fig. 70).

The following rule, therefore, may be laid down for the summer season: Whereever in the gulf the surface temperature is lower than in the surrounding waters, and the water is nearly homogeneous vertically (with little stability), vertical circulation of some sort is active; but where the water is warmest at the surface and most stratified as to temperature and density vertical circulation of any kind is weakest.

Nantucket Shoals, where tides run strong over and between the ridges and channels, afford an interesting example of the thermal result of active mixing, the surface temperatures there being lower but the bottom water warmer in summer than in the neighborhood generally. These same criteria show active mixing on the eastern side of Georges Bank; likewise, no doubt, about Georges and Cultivator Shoals. This also applies to German Bank. Dawson (1905, p. 15) has pointed out that pools or "wakes" of low surface temperature, extending north and south from Lurcher Shoal off Yarmouth, result in the same way "from the stirring up of the water."

The Bay of Fundy is the classic example of violent tidal stirring for the Gulf of Maine region, where the currents, running with great velocity over the shoals at its entrance and among the islands off the New Brunswick shore, keep the water mixed, top to bottom, throughout the summer, a fact referred to repeatedly in the preceding pages. This peculiar circulatory state of the Bay of Fundy, made clear by Huntsman, is of far-reaching biologic significance; for, as he points out, so low a surface temperature is thereby maintained throughout the summer that "conditions approximating those in the far North are produced in shallow water" (Huntsman, 1924, p. 281).

The rush of the tides between the islands along the coast of Maine, east of Penobscot Bay, is similarly reflected in low stability and slight thermal stratification (p. 599).

[87] Only about one-third as stable near Mount Desert and one-tenth as stable near Grand Manan as at the mouth of Massachusetts Bay.

The courses of the curves for 1.5, 2, and 2.5 units of density on the chart (fig. 186) give evidence that the shoal ground off Penobscot Bay and out to Cashes Ledge also is the site of considerable vertical disturbance as the tidal currents are deflected by it.

As summer passes into autumn and the surface waters commence to cool, the parts of the gulf that are most stable in summer become less and less so, with little change in the eastern part, where the whole column of water loses heat more uniformly. The result is that vertical mixing is less and less opposed in the western part of the gulf and regional differences decrease in this respect.

The autumnal decrease in stability is illustrated for the southwestern part of the gulf, generally, by the offing of Cape Ann, where the upper 40 to 50 meters lose stability most rapidly during the early autumn, then more slowly but constantly through the winter. At depths greater than 100 meters no regular seasonal succession appears, all the curves being roughly parallel, their differences attributable to annual fluctuations in temperature and salinity. The seasonal succession is essentially of this same kind in the deep water in the northeastern corner of the gulf, though, thanks to strong tidal currents, the seasonal range of stability in the upper 40 meters (expressed in terms of density) is only about one-third as wide here as it is off Cape Ann.

Stability offers but little opposition to the free vertical circulation of water in any part of the gulf after November; less near the surface than at greater depths, as appears from the following table for October and November of 1916:

Vertical range in density for the superficial stratum and for the mid stratum

Station	0 to 40 meters	40 to 100 meters	Station	0 to 40 meters	40 to 100 meters
10399	0.79	1.00	10402	0.12	1.00
10400	.54	.90	10403	.55	1.30
10401	.51	1.35			

The free mixing that takes place from that time on throughout the winter is illustrated by the uniformity with which the upper 50 to 100 meters cool off during December, January, and the first half of February; evidently, water is constantly being brought up to the surface from below, there to radiate its heat, and water cooled at the surface is as constantly sinking.

It is not necessary to follow in detail the changes in stability that take place in winter in this connection. It is lowest over the gulf as a whole at the end of February or first of March, when the difference in density between the surface and the 40-meter level has been only 0.1 to 0.33 for all our stations on the banks and within the gulf, except at one off the Kennebec River (station 20058).

In fresh-water lakes, in high latitudes, autumnal cooling increases the density of the surface until a dynamic overturning of the water regularly follows. Our first winter's work in Massachusetts Bay (Bigelow, 1914a, p. 387) suggested that this same process was partly responsible for the rapid chilling that takes place there; but subsequent study, and especially the observations made in the bay from the *Fish Hawk* in 1925, proves this earlier interpretation erroneous and make it unlikely that dynamic overturning ever occurs in the open gulf, unless on a small scale and confined to a very thin superficial stratum. This statement is based on the fact that the density has been slightly lowest at the surface at all our winter stations, when compression is allowed for, though without this factor the surface stratum would often appear heaviest. It is true that the stability of the water is virtually *nil* in winter; but tidal stirring and the stirring effect of the wind are everywhere so active during the cold months that they more than keep pace with the chilling of the surface by constantly bringing up new water from below to take the place of the surface layer as the latter chills and before it is heavy enough to sink.

The thermal effect of mechanical mixing is essentially the same as that of dynamic overturning, however—i. e., to bring the whole column of water within the chilling influence of the low air temperatures. It is possible that dynamic overturning does occur locally in the coastal zone, but it has not actually been recorded there.

Vertical dynamic circulation of another sort was observed in Massachusetts Bay in February, 1925, where water, chilled at the surface close to the land, was moving offshore on the bottom, and with surface water from offshore moving in above it to take its place, as described above (p. 659). A more detailed survey of the temperature of the coastal belt in winter may show that circulation of this sort is more widespread than appears from observations taken so far.

DYNAMIC EVIDENCES OF CIRCULATION

CONSTRUCTION OF DYNAMIC CHARTS

Given a difference of pressure between any two stations in the sea, a current will result as surely as water will flow out through a dam when the sluice gate is opened, unless opposed by a stronger counterforce or an unpassable barrier. Even a preliminary examination of the dynamics of the gulf (and no more is attempted here) may be expected greatly to amplify such knowledge of its dominant circulation as has been gained from the more direct lines of evidence discussed in the preceding chapters.

The method of attack chosen here is that of the dynamic-contour chart, widely employed by European oceanographers and recently described by Smith (1926). For the sake of the nontechnical reader, an explanation of the principles involved in its construction and its interpretation are attempted here in the simplest possible language.[88]

In the sea, gravity, acting always directly downward, will set the water in motion if its surface slopes at all; and even if the surface of the water be level, currents will be caused if its specific gravity is greater at one place than at another, because the pressure exerted by the water at a given depth must then vary correspondingly, and the plane at which the pressure is uniform must be oblique to the pull of gravity. All this is embodied in the old adage, "water seeks its own level."

Although the physical principles that govern the gradient currents in the sea are simple, calculation of the drifts that will actually result from any given distribution of specific gravity is so complex that Bjerknes's (1898, 1910, and 1911) illumi-

[88] See also Sandström (1919) for a simple explanation of hydrodynamic principles.

nating application of mathematical methods first offered a practical and easy method of solution.

Since that time European, and especially the Scandinavian, oceanographers have devoted much attention to the dynamic calculation of ocean currents, with such success that great advances in our knowledge of oceanic circulation are to be expected. Sandström (1919) has also studied the dynamics of Canadian Atlantic waters; Wüst (1924) of the straits of Florida and neighboring parts of the Atlantic; and Smith (1926, 1927) of the "Labrador" and "Gulf Stream" currents around the Grand Banks.

The simplest and most graphic method of learning the directions followed by the dynamic circulation in any sea area is by a horizontal projection showing (by contour lines) the regional variations in the thickness of the column of water included between the surface of the sea and the level at which some given pressure, equal for the whole area, is reached.

If the specific gravity[89] of the water is regionally uniform over the whole area, the depth of the layer so bounded will equally be uniform, and there will be no dynamic flow from any one part of the picture to any other; but if the weight of an equal thickness of water be greater (i. e., its specific gravity higher) at one locality than at another, a lesser thickness will produce a given pressure at the heavy station rather than at the light, and such a flow will tend to develop.

Consequently, calculation of the height of the column of water necessary to exert a given pressure for any two stations will give the dynamic tendency existing between them in the stratum included in the calculation; and if the survey can be extended to include a number of stations, scattered netlike over any part of the sea, we arrive at the dynamic gradients for the whole area.

This calculation is based on the principal that the pressure exerted by a column of water of unit area is the product of three arguments—its height, its specific gravity, and the acceleration of gravity; and if the first and the last of these be combined into dynamic units of measurements, as explained below (p. 932), pressure may be stated still more simply as equal to the height of the column (in dynamic units), multiplied by its specific gravity. Or, conversely, the height of the column (in dynamic units) will equal the pressure it exerts, multiplied by the reciprocal of the specific gravity of the water, namely, by its specific volume.

For example, if the specific gravity of a given column of water be 1.026, and it be desired to find the height or depth (in dynamic units) necessary to exert 50 units of pressure, we have: Specific volume $0.97466 \times 50 = 48.73300$ dynamic units of depth. If at a neighboring station the specific gravity is only 1.022, 48.92350 units of depth will be requisite to effect this same pressure, so that there will be a dynamic slope between the two stations of 0.2 dynamic units of height (or depth).

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[89] A brief definition of the much-abused term "density" as employed to express the specific gravity of sea water follows:

In hydrodynamic calculation what is important is the specific gravity that the water in question actually possessed at its temperature at the time and under the pressure to which it was actually subjected—i. e., *in situ;* not that which it might have possessed at any other temperature or depth.

The specific gravity of sea water differs from that of distilled water only in the second and subsequent decimal places. To avoid the use of such long decimal fractions it is usual to subtract 1 and to multiply by 1,000, substituting the term "density" for "specific gravity." For example, the density of sea water of a specific gravity of 1.025 is stated as 25.00.

Specific volume (merely the reciprocal of density) is the more convenient value to use in numerical calculation.

The practical application of this theorem to hydrographic problems thus hinges on the selection of suitable unit values for thickness and for pressure; the selection of such was not the least of Bjerknes's contributions to dynamic oceanography.

The force responsible for dynamic currents in the sea is that of gravity—not the capacity for work inherent in the water itself because of its mass. Consequently, the unit of height (or thickness) used in hydrodynamic calculations must not only stand in a linear relationship to the unit of pressure, but it must also be a direct measure of the potential force of gravity, which accelerates all falling bodies equally, irrespective of their mass. The gravity potential set free when a unit mass of water flows down a sloping surface is the product of two arguments—(1) the vertical difference in height and (2) the accelerating force of gravity. The latter being about 9.8 meters per second, the dynamic value of 1 meter of linear height must (in the meter-ton-second system) be stated as 9.8 units. Thus, gravity performs one unit of work in $\frac{1}{9.8}=0.102$ meters, so that one dynamic decimeter = 0.102 meters, or one dynamic meter = 1.02 common meters. For the reason just stated this relationship between dynamic and common linear measure is constant, no matter what the density of the water under study may be.

It is not practical to make direct instrumental measurement of the pressure below the surface of the sea; this can be deduced only from measurements of the temperature and salinity, and these must be taken at predetermined depths.

To calculate the thickness of a column of water that will exert any given pressure—say 100 units—the first step then is to establish the specific volume. This decreases in the sea with depth; consequently, to learn the mean specific volume it is necessary to determine the value not only for the top but also at the bottom of the column. If we could know before hand how deep it would be necessary to lower our instruments in order to do this—in other words, if the pressure unit of thickness could correspond to the ordinary linear measure—evidently the procedure would be vastly simplified. Strictly speaking, this is impossible because the linear value of this pressure unit *must* vary with the specific volume of the water. In practice, however, as Bjerknes and Sandström and Helland-Hansen (1903) have explained, this objection vanishes because the specific volume of the water varies only so very slightly with depth that the value will be given for the bottom of the chosen pressure column if the readings are taken within a few meters of it, whether shoaler or deeper.

Consequently, if a pressure unit can be found, which shall nearly (even if not quite) correspond to the ordinary linear measure, we can learn the specific volume where the pressure is, say, 100 units, simply by measuring the specific volume at a depth of 100 meters. The selection of such a unit we owe to Bjerknes, who proposed the "bar" to be equal to the pressure exerted by 10 dynamic meters (or 10.2 common meters) of fresh water, not under compression, and at the temperature of its maximum density. By the theorem stated on page 931, that pressure is the product of linear height, specific gravity, and acceleration of gravity, the "bar" will then equal 9.9 meters of salt water 35 per mille in salinity and 0° in temperature, so that a decibar is virtually 1 meter of sea water. For the reasons just stated, if the salinity and temperature be taken at any chosen number of meters below the surface this will give the specific volume where the pressure is that same number of decibars. Thus, if in the example given on page 931 we read dynamic meters instead of units of thickness, the corresponding units of pressure will be 50 decibars.

If the dynamic depth to which it is necessary to descend into the sea to reach a given pressure be greater at one station than at another (as is necessarily the case if the specific gravity of the water varies regionally), only two alternative states are possible: (1) If the surface of the water is level, the given isobaric surface (surface at which the pressure is equal) must slope; or (2), if this isobaric surface is level, the surface of the sea must slope. The resultant circulation will differ accordingly.

If the first alternative actually prevailed, the obliquity of the isobaric surfaces would increase with depth and the dynamic circulation would be most rapid at the bottoms of the deepest oceans. However, as Sandström (1919) and Smith (1926) both have emphasized, this is directly contrary to the truth, for the bottom waters of the ocean show only very slight regional variations in specific gravity and move only with inconceivable slowness. Consequently, when a dynamic gradient exists over any part of the sea it is the surface that slopes. It is of the greatest importance to keep this concept constantly in mind, because the conventional dynamic representations in profile show the surface as level, and hence are likely to prove misleading.

If, then, the isobaric plane chosen as the base for reference in our calculations lies so deep that it is level, or virtually so, calculation of the thickness of the column of water necessary to effect this pressure for a number of stations shows the actual contour or shape of the surface of the sea. Dynamic-contour charts of the deep oceans, such as have been constructed by Helland-Hansen and Nansen (1926) and by Smith (1926), are cases in point. In shoaler waters, however, where surfaces of equal specific gravity, and consequently the isobaric surfaces, are oblique right down to the bottom, the calculated dynamic slope of the surface of the sea will either exaggerate or minimize the true slope of the latter.

This is the case in the Gulf of Maine. Consequently, the dynamic charts offered here can be taken only as a rough approximation to the state actually prevailing.

The actual charting of the dynamic gradients in horizontal projection is hardly as simple as the foregoing résumé might suggest because of the necessity for intergrating the individual values for specific gravity at the levels of observation to arrive at the mean values for the included intervals; because, also, the specific gravities must be converted into specific volumes, and because the latter must be corrected for compression. The last two steps, however, are robbed of all difficulty by Hesselberg and Sverdrup's (1915) tables, as simplified by Smith (1926, p. 18, Tables 3 and 4). Smith (1926) has so fully explained the construction of the dynamic chart, as well as the principles involved, in a publication universally accessible, that only one aspect of the procedure needs further comment here, namely, the modifications necessary in studying an area so shoal and with stations differing so widely in depth that it is not possible to refer all the calculations to any one isobaric base plane. In this case it is necessary to calculate the gradient between pairs of adjacent stations, afterwards referring all to some one chosen station. Furthermore, if the specific volumes of the water at the two members of each pair of stations are not the same at the greatest depth reached at the shoaler, it is obvious that the intervening mass of bottom water deeper than that level must be in dynamic circulation;

hence, it must be taken into account in some way in calculating the dynamic slope at the surface.

Jacobsen and Jensen (1926) have very fully discussed this question in their dynamic study of the Faroe Channel, finding that in most cases this effect of the bottom water may be sufficiently allowed for by arbitrarily applying to the dynamic gradient between the two stations in question the product of the difference in specific volume between them at the deepest level of the shoaler station multiplied by half the difference in depth. If the station where the calculation shows the surface as highest also has the largest specific volume at the deepest level of the shoaler of the pair, the gradient is to be increased by the amount of this correction—decreased if the reverse obtains. If the difference in depth be greater than, say, 150 meters or so, no arbitrary correction of this sort can be relied upon, consequently the dynamic gradient can be stated only within very wide limits. The only cure is to establish the stations closer together on future cruises.

The dynamic-contour chart[90] closely resembles an ordinary weather map in its general appearance, and it is as easily interpreted in terms of the resultant circulation. Dynamically, the water tends to flow down the slopes from the parts of the picture where the surface stands high to those where it is low, and at right angles to the contour lines. Actually, however, this could happen only at the equator. Everywhere else the effect of the earth's rotation so deflects this motion that the stream lines come nearly to parallel the contour lines, which may then be taken as directly representing the current, just as the direction of the wind is roughly parallel to the isobars on the weather map.

In the open ocean, where tidal currents are weak, the contour lines may even approximate the tracts of the particles of water if approximately constant acceleration has been established. This, however, does not apply in a region such as the Gulf of Maine, where the tidal currents average much stronger than the dynamic tendencies. In this case the latter act only to give to the tidal flow a character more definitely rotary than would otherwise be the case, or to strengthen the one tide at the expense of the other. Here the dynamic-contour lines show only the general advance which the water tends to make good in its tidal oscillations to and fro.

Because in every case the datum plane for the calculation is necessarily the underlying water, not the solid bottom of the sea, the motion indicated by the chart is not absolute, but is only relative to that of the deepest stratum of water included in the picture. If this be motionless, the calculated drift represents the actual motion of the surface (or chosen level) relative to the coast line, but not otherwise.

In the Northern Hemisphere, where moving bodies are deflected to the right, the direction of flow, relative to the plane of reference,[91] is to be identified by the rule that the gradient current will constantly have the lightest water (i. e., the highest surface) on its right hand, the lowest surface on its left, as it veers cyclonically around the latter. If the surface drift be faster than the bottom drift, as is usually the case, this indicated direction of flow will also be the true direction, relative to the bottom; so, too, if bottom and surface drifts be parallel, whichever

[90] Dynamic-contour charts may as easily be constructed for any desired depth below the surface of the sea, as described by Smith (1926).

[91] In the Gulf of Maine this is the bottom water between the pairs of adjacent stations.

is the stronger. But if the bottom current be the stronger, and both currents are opposite or diverge by a considerable angle (as may rarely be the case in shoal water, though perhaps never in deep), the method is made unreliable.

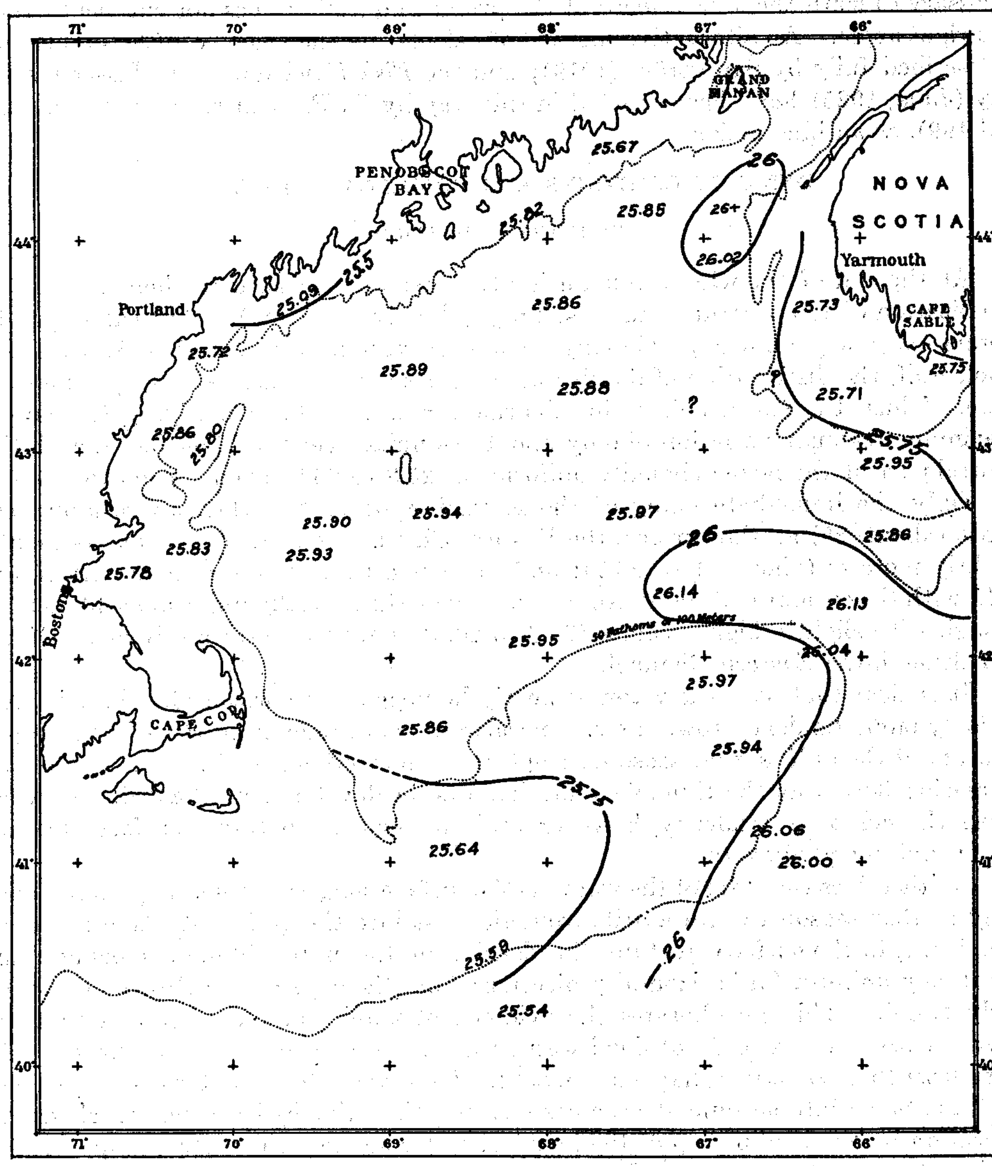

FIG. 187.—Distribution of density at the surface, February to March, 1920

In studying the dynamics of any shoal area it is also essential to appreciate the effect which the contour of the bottom may have in deflecting the gradient currents. This, of course, can not be stated by rule, but usually it is fairly simple of interpretation.

Once the dynamic gradient is established between any two stations, the corresponding velocity of the water at the one, relative to the other, is calculable by a simple formula, described by Smith (1926, p. 31), who also makes clear the correction necessary to learn the true velocity if the profile in question does not cut the current at a right angle. An alternative method of calculating the velocity, often employed, is described fully by Sandström (1919); and the *Fish Hawk* data for Massachusetts Bay (June, 1925) have been treated in this way by R. Parmenter (p. 949; figs. 198 and 199), as an illustration.

DYNAMIC CONTOURS AND GRADIENT CURRENTS

FEBRUARY AND MARCH

At the end of the winter and during the first days of spring, when the general equalization of temperature and of salinity (already discussed) makes the upper 40 meters extremely uniform, regionally as well as vertically (pp. 522, 703), over the whole gulf, the distribution of density at the surface would suggest a very quiescent state. Thus, the surface chart for February and March, 1920 (fig. 187), shows a maximum regional variation of only about 0.4 units over the whole basin, with the central part of the latter virtually uniform (at 25.8 to 25.9) from station to station.

Only the immediate offing of the Kennebec River was then appreciably less dense (about 25) at the surface, the Eastern Channel and the region off its mouth slightly more so (about 26 to 26.1); and the whole western and central part of the gulf, with the coastal belt along Nova Scotia, was then equally uniform at 40 meters, though with slightly higher values (26.3 to 26.5) along the eastern side of the basin and through the Eastern Channel.

It is clear that with the water so nearly homogeneous horizontally there is very little dynamic tendency toward any general system of gradient currents in the upper stratum of the gulf at that season, except that the freshening of the surface by the increasing flow from the Kennebec foreshadows the development of a drift westward along the coast—a tendency, however, still confined to so thin a surface stratum that it did not yet govern.

Neither does the state of the water at the surface suggest a general dynamic tendency at that season toward a drift from the east into the gulf past Cape Sable, or vice versa, in the surface stratum, the density of the upper 40 meters being comparatively uniform (in horizontal projection) from the cape out to Browns Bank for early March. This corroborates the evidence of salinity and temperature that the Nova Scotian current did not flood westward past the cape in the spring of 1920 until later than sometimes happens (p. 832). However, when the density of the deep strata is taken into account it becomes obvious that the hydrostatic forces set in operation by the banking up of the heaviest water against the eastern slope of the gulf (p. 849, fig. 172) must tend to cause a cyclonal or anticlockwise movement of the deeper mid strata, carrying with it, as an overlying blanket, the surface stratum, itself so nearly quiescent.

The dynamic chart for February and March, 1920 (fig. 188), gives an indication of the stream lines to be expected at the surface under the conditions of temperature and salinity then existing, which may be taken as typical of the first two weeks of spring. However, I must here caution the reader that at this time of year, when the propulsive force for gradient currents is derived mostly from the deep strata of

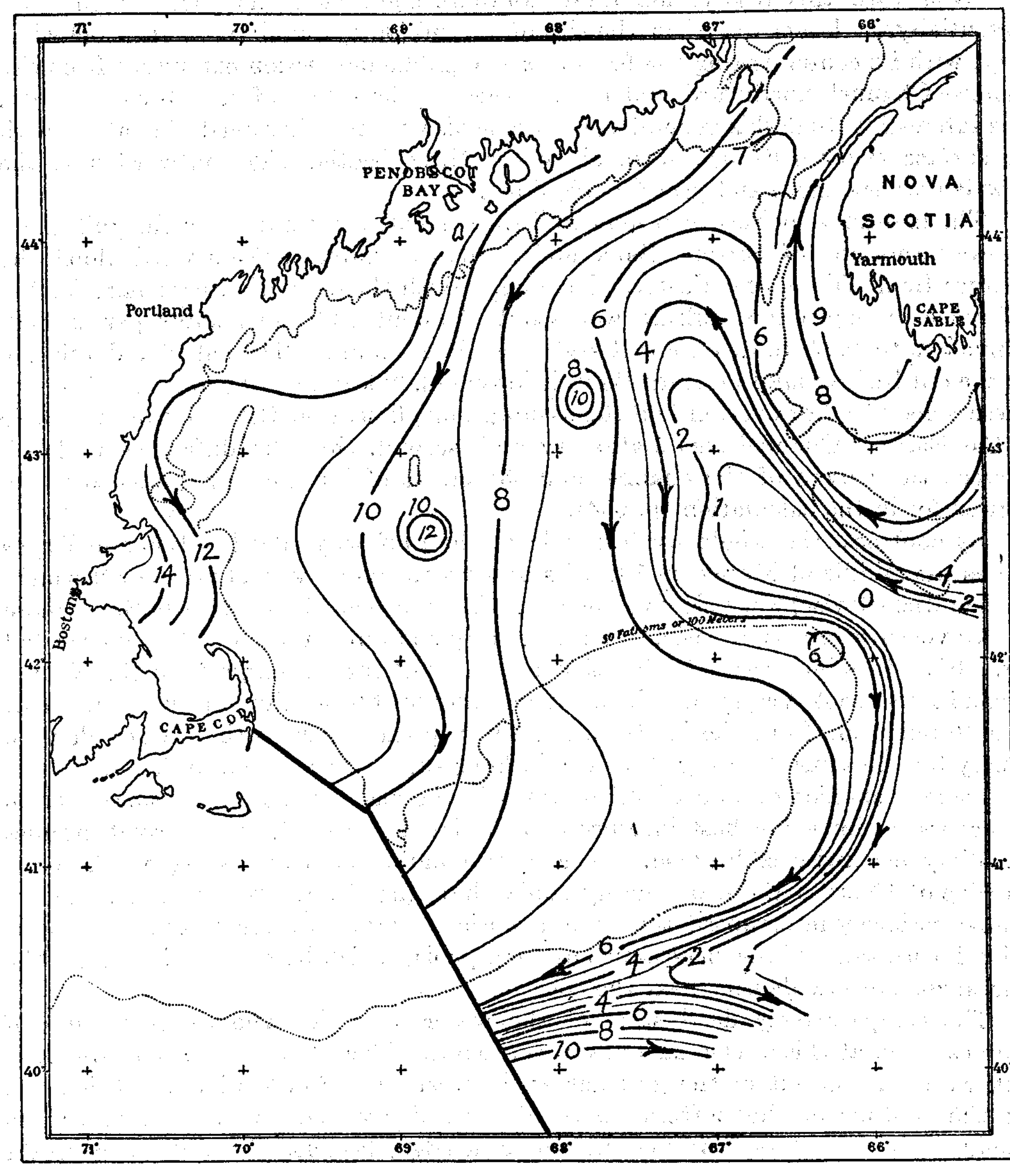

FIG. 188.—Dynamic gradient at the surface of the gulf for February to March, 1920, referred to the Eastern Channel as the base station. The dynamic heights are given for every dynamic centimeter. For further explanation see p. 937

water, the probable error introduced into the calculations by the necessity for assuming an arbitrary correction for the differences in depth between pairs of adjacent stations (p. 934) is relatively greater than for late spring or summer, when the surface stratum is moving more rapidly than the underlying water. Consequently, the contour lines on the early spring chart (fig. 188) and the dynamic gradients which they show can be accepted only as a rough approximation, not in detail. Some smoothing of the curves has proved necessary in the construction of the chart, also.

Even with this reservation these contours show that the basin of the gulf (potentially, at least) was then the site of one major cyclonal (i. e., anticlockwise) eddy, with its center taking the form of a troughlike depression extending from the Eastern Channel northward and inward toward the offing of the Bay of Fundy. It is interesting that this general eddy seems also to have involved the latter, with the surface water drifting inward along the Nova Scotian side, outward next the New Brunswick shore and past Grand Manan.

The highest velocities then indicated were a drift northward into the gulf along the western slope of Browns Bank and a counter movement outward along the Georges Bank side of the Eastern Channel. With the correction used here for the difference in depth this indraft works out at about 13.5 centimeters per second, equivalent to 0.27 knot, or about 6½ miles in 24 hours. The calculated velocity for the outdraft around Georges Bank is lower—0.22 knot, or 5¼ miles in 24 hours. These velocities, however, are on the assumptions, first, that the water in the center of the Eastern Channel was stationary and, second, that the difference in depth between the trough of the channel and the crests of its two slopes was correctly allowed for in the calculation (p. 934).

By contrast, the whole western side of the gulf was "dead," dynamically, as late as the middle of March, in 1920, its upper stratum only tending to drift southward (anticlockwise) very slowly, except at the mouth of Massachusetts Bay, where greater velocity in this direction is suggested by contour lines more closely crowded (fig. 188). It is interesting to find that the effect of the discharge from the Kennebec and Penobscot was most evident in speeding up the southwesterly surface drift some 40 miles out from the land—not close in to the latter, as the surface chart of density for the same date (fig. 187) would have suggested if taken by itself.

Lower densities at two of the stations in the basin (20054 and 20052) than in the general vicinity are best interpreted as isolated pools, which, if correct, implies sudsidiary clockwise eddies; so, too, a corresponding high appearing on the eastern edge of Georges Bank on the dynamic chart (fig. 188). While these seem not to have seriously interrupted the general anticlockwise movement, they are interesting illustrations of the persistence of such pools, which have drifted off from the general zone of low density next the coast.

The comparatively dead state of the water over the whole eastern half of Georges Bank at this season also deserves a word. The chart suggests a slow drift southward and so out of the gulf across the western half of the bank at this time, but the contour of the bottom makes it more likely that the surface water was actually moving eastward around its northern edge, because the underlying strata (which in this case supplied the motive power) are necessarily directed by the submarine slope, against which any southward drift must strike. Thus, we may conclude that the dynamic movement of water around the basin was even more definitely eddylike and anticlockwise in March than the chart (fig. 188) suggests.

Lacking March data for the region of Nantucket Shoals, the chart fails to show whether a definite dynamic outflow is to be expected around the latter to the westward from the gulf at that season.

In the offing of Cape Sable the dynamic gradient for March,1920, calls for a weak drift clockwise but spreading far offshore toward Browns Bank before eddying northward again toward the gulf. Hence, thecold Nova Scotian water that we encountered midway out over the shelf (station 20075, p. 1000) did not then tend to round the cape, but to veer offshore, which agrees with the distribution of temperature and salinity at the time. Dynamic evidence also is strong that whatever water was then entering the eastern side of the gulf in the upper stratum was drawn chiefly from the region of Browns Bank and from the edge of the continent in the offing of Cape Sable—i. e., from the source whence the gulf regularly receives its slope water (p. 848).

The dynamic gradients for March are especially instructive along the continental slope abreast of the gulf because of the light they may throw on the problem of the so-called "Gulf Stream" along this sector. Fortunately, this is made comparatively clear for this region (fig. 188) by the considerable difference in density between the outer stations on the two cross profiles of the bank—western and eastern (stations 20044 and 20069). On the eastern profile the gradient (dipping to a low at the outermost station) shows a strong drift to the westward along the edge of the bank, its calculated velocity being about 0.6 knot, or 14 miles in 24 hours. While this calculation depends on the correct allowance for the difference in depth between stations, one of which was much deeper than the other,[92] the direction of this gradient current is well established. A weak continuation of this westerly drift (indicated by a low in the dynamic contour) extended along the edge of the bank as far as the western profile (run three weeks earlier); but here this gave place to a much steeper counter gradient to high in the next 10 miles offshore, implying a counter drift to the east.

Unfortunately, the difference in depth between the stations on the edge of the bank and outside is again so great on this profile (150 to 200 and 1,000 meters) that the arbitrary correction employed to take account of it becomes only a rough approximation, though the order of this correction (i. e., whether increasing, decreasing, or even tending to reverse the gradient calculated for equal depths) is in every case clear enough (p. 934). When all reasonable allowance is made for this source of error, however, the velocity of the easterly drift may safely be set as at least half a knot. Fortunately, calculation of the dynamic head between the two outermost stations on these two profiles is not subject to this error, both being deep enough (1,000 meters) to reach equal density at the lowest levels. Consequently the general contour, as laid down for this region in Figure 188, is established, as is the fact that the western profile reached out to water of comparatively high temperature and salinity in the upper stratum, while the eastern profile did not, though its outermost station was still farther out from the edge of the continent.

So long as the dynamic gradient continues to be of this sort it is evident that the superficial drift of warm water along the continental slope, commonly spoken of as the "inner edge of the Gulf Stream," is not only to be described as a typical gradient current but is to be expected within 15 to 20 miles of the edge of the bank between longitudes 68° and 69°. Farther east, however, the contour lines on the chart (fig. 188) show it departing farther and farther from the bank, agreeing in this

[92] Station 20068, 200 meters; station 20069, 1,000 meters.

with general report. On the other hand, the westerly counterdrift set in motion along the inshore side of the dynamic depression (or cabelling zone) loses in velocity and hugs the bank more closely from east to west.

From the general oceanographic standpoint this demonstration that this sector of the "Gulf Stream" receives a propulsive impulse from the local hydrostatic forces (i. e., is strictly a dynamic drift) is one of the most interesting results of our explorations.

The upper 50 meters or so of the gulf being close to quiescent, dynamically, during February and March, the chart for the surface (fig. 188) will as well represent the gradient currents down to as deep as 100 meters or so for that season, leading to the interesting result that the whole column down to this depth tended to drift inward along the eastern side of the Eastern Channel at the time, outward along its

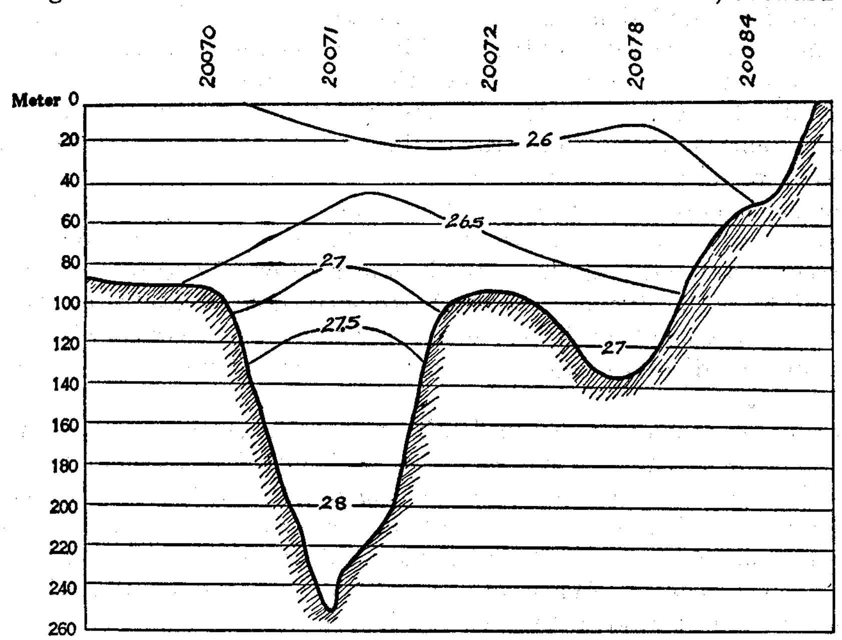

FIG. 189.—Distribution of density on a profile running from the eastern end of Georges Bank across the Eastern Channel, Browns Bank, and the Northern Channel, to the vicinity of Cape Sable, March 13 to 23, 1920. Corrected for compression.

western side, which is also evident in the profile (fig. 189). However, if we descend to as great a depth as 150 meters a rather different dynamic distribution appears, with the center of anticlockwise revolution located as a low close to the northern slope of Georges Bank, with a weak but definite tendency toward a gradient drift crossing the basin from northeast to southwest, shown better graphically by the dynamic contours (fig. 190) than verbally. This drift was then bounded on the west by a considerable dead area covering the whole west-central part of the basin (except as interrupted by a subsidiary high marking a clockwise whirl in the offing of Penobscot Bay), with a very weak southerly tendency along the western slope in the offing of Massachusetts Bay.

In the eastern side of the area this deep projection points to a slow creep inward through the Eastern Channel; but with only one station in the latter it is impossible to state whether this creep involved the whole breadth at this depth or (which seems more likely) hugged its Browns Bank slope, as in the shoaler strata.

In interpreting the dynamic contours in terms of potential drift at a depth at which the basin of the gulf is entirely inclosed except for one narrow channel, it is obvious that prime consideration must be given to the contour of the bottom, as this controls the possible movement of the water. When this is taken into account, the March chart (fig. 190) affords the best clue yet available to the movement of the slope water over the floor of the gulf at a season when this is entering in large volume via the trough of the Eastern Channel (p. 850). Dynamic contours for the 150-decibar

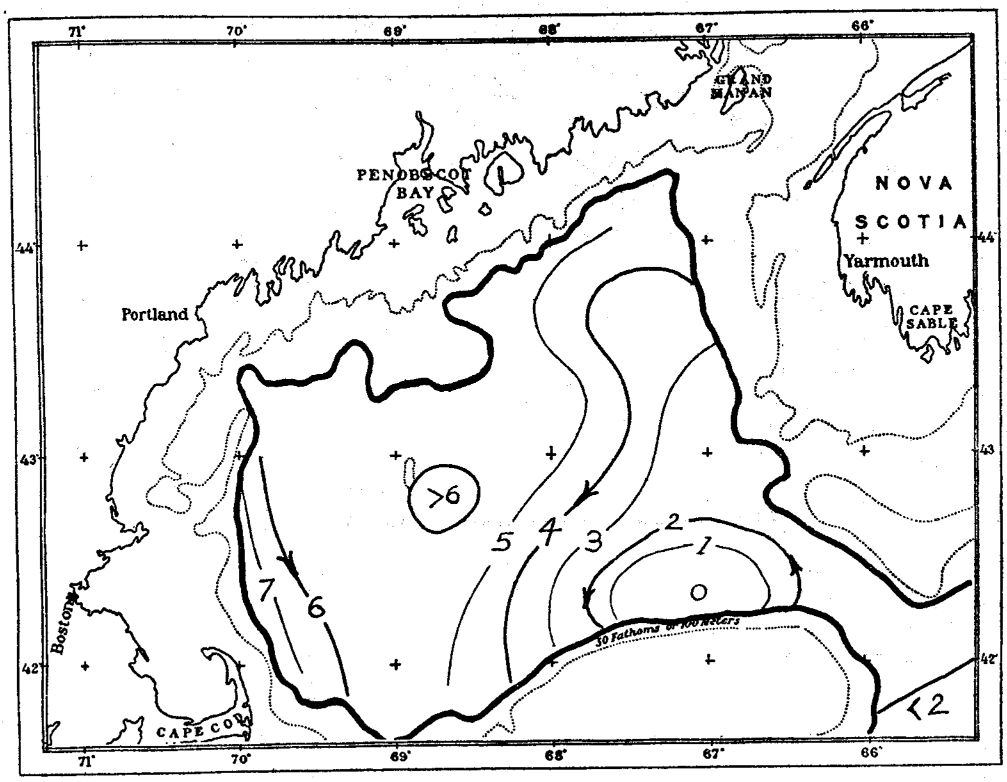

FIG. 190.—Dynamic gradient, bottom to 150 decibars, referred to the southeastern side of the gulf as base station, for February and March, 1920. Contour lines for every dynamic centimeter

level, like the distribution of temperature and of salinity, show this indraft following the eastern side of the basin inward, to eddy westward and so southward; but instead of completing a circuit around the cyclonic center ("low" on the chart—fig. 190), the drift will obviously be deflected by the slope of Georges Bank. The angle at which the contour (or stream) lines strike the latter suggests an overflow into the dead western side of the basin. It is here, then, as well as along the northern slopes of the gulf, that the consumption of this slope water chiefly takes place during the early spring, as tides and wind currents constantly mix it with the less saline but colder stratum above.

The implication of a dynamic contour of this sort in the deeps of the gulf, combined with the effect of the confining slopes and with this consumption in the inner part, is obvious; it provides a propulsive force to pump into the gulf the slope water with which the offing of the Eastern Channel is kept supplied—also dynamically—from the source of manufacture to the eastward (p. 847).

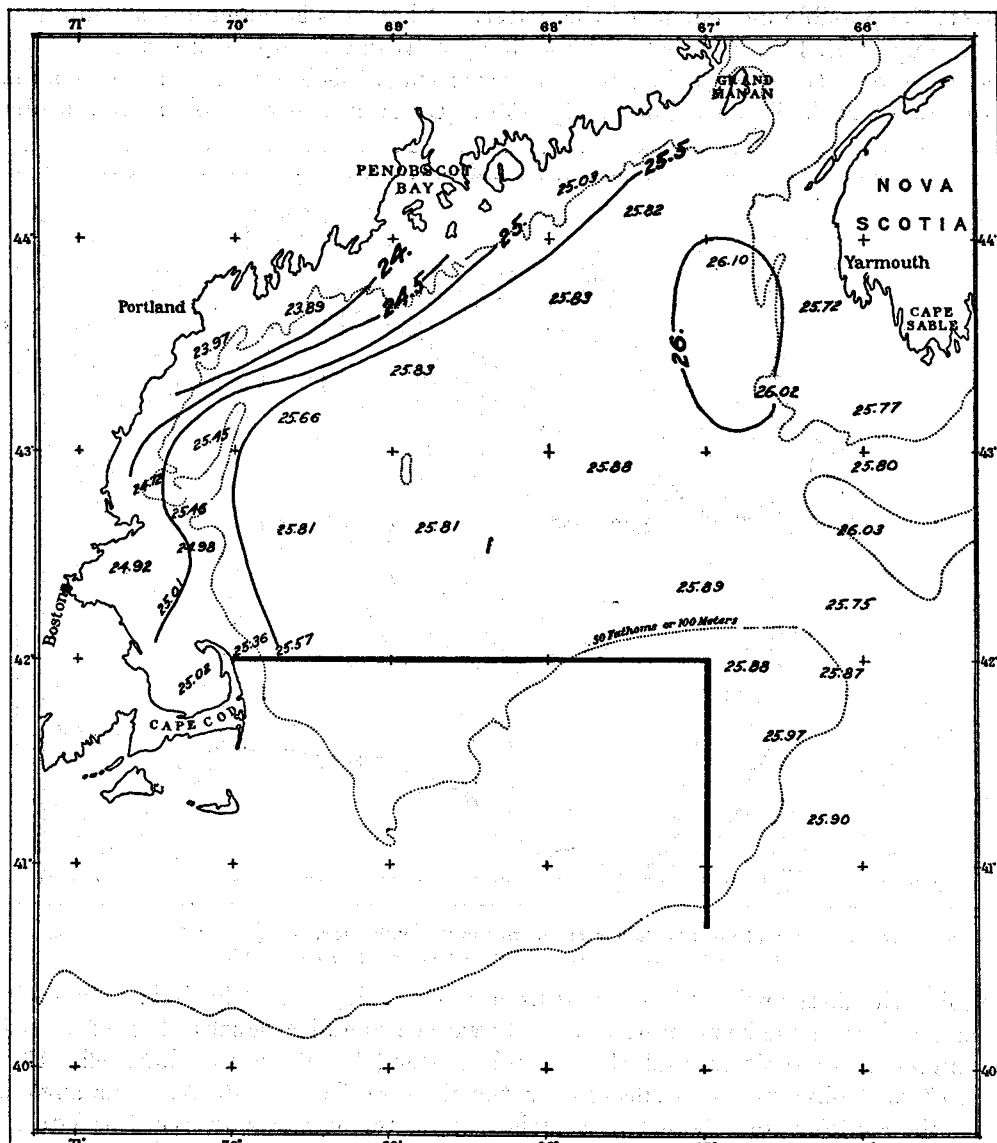

FIG. 191.—Distribution of density at the surface of the gulf, April 6 to 20, 1920

APRIL

The progressive freshening of the surface, which takes place along the northern and western shores of the gulf with the advance of spring, results in the development of a corresponding coastwise belt of low-surface density by April, grading abruptly to considerably higher values a few miles out in the basin (fig. 191). This development adds both velocity and volume to the longshore drift west and south, which was foreshadowed on the March chart (fig. 188).

In 1920, according to the dynamic contours at the surface (fig. 192), this spring current had come to dominate the entire coastal belt of the gulf from the neighborhood of Mount Desert Island (probably from the Grand Manan Channel) to Cape Cod by the middle of April, and probably it does so every year by this date—earlier in years when vernal progression in the sea is more forward. During the period covered by this April cruise the average calculated rate of this current, referred to the "low" in the offing of the Bay of Fundy (assumed stationary), was about 0.3 knot abreast of Mount Desert, about 0.18 knot abreast of Cape Cod, or an average drift of about 5¾ miles per 24 hours along this coast sector as a whole. In spite of the sources of unavoidable error this calculation falls at least within the order of magnitudes suggested by other lines of evidence.

In Massachusetts Bay, also, a continuation of this longshore drift is indicated by the dynamic contours from the north shore around toward Cape Cod. This, again, agrees with the drifts of bottles that were set out a few miles north of Cape Ann in April, 1925 (p. 890; fig. 177); and evidently this is the characteristic state during that month, for salinities and temperatures taken in the bay by the *Fish Hawk* on April 21 to 23, 1925, show a drift of low density (fig. 193) southward past Cape Ann and across the mouth of the bay to Cape Cod as the water from the Merrimac and other rivers to the north floods southward.

Surface projection (fig. 191) and dynamic contours (fig. 192) for April unite in locating the low in the offing of the Bay of Fundy some 60 miles off Mount Desert Island for that month, the whole east-central part of the basin out through the Eastern Channel being virtually dead dynamically, contrasting with a weak northerly set along the western shores of Nova Scotia. In the southern side of the area the dynamic contours point to a persistence of the drift out of the gulf to the south around the eastern end of Georges Bank, just described for March (p. 938; fig. 188), though at a lower velocity; but as a result of the equalization of temperature and salinity from the Eastern Channel in across Browns Bank (p. 553) only a very slow movement into the gulf along this side of the channel is suggested by the April chart (fig. 192).

The general result of the lightening of the northern and western margins of the gulf, combined with the shift of the cyclonal low northward across the basin, which follows a slackening in the indraft of slope water, is to give the anticlockwise circulation more definitely the character of a great eddy in April than in March, centering off the Bay of Fundy and with its western side traveling southward with greater velocity than its eastern side drifts north.

It is probable that in April the gradient currents are given an easterly direction along the northern slopes of Georges Bank, just as in March (p. 938), by the contour of the bottom, with a separation off Cape Cod between this easterly drift and a southerly drift past the cape and past Nantucket Shoals. This suggestion is corroborated by the fact that bottles followed both these routes from Massachusetts and Ipswich Bays in April, 1925.

MAY

Progressive incorporation of river water into the northern and western sides of the gulf, coupled with vernal warming, constantly favors the anticlockwise movement of the so-called "spring current" (fig. 194); and with the resultant changes in salinity and temperature affecting chiefly the surface, the site of the chief dynamic impulse toward circulation shifts from the deep strata to the superficial. In May, 1915, for example, a difference of about 1.5 units of density was recorded at the surface between the vicinity of the mouth of Massachusetts Bay and the basin in

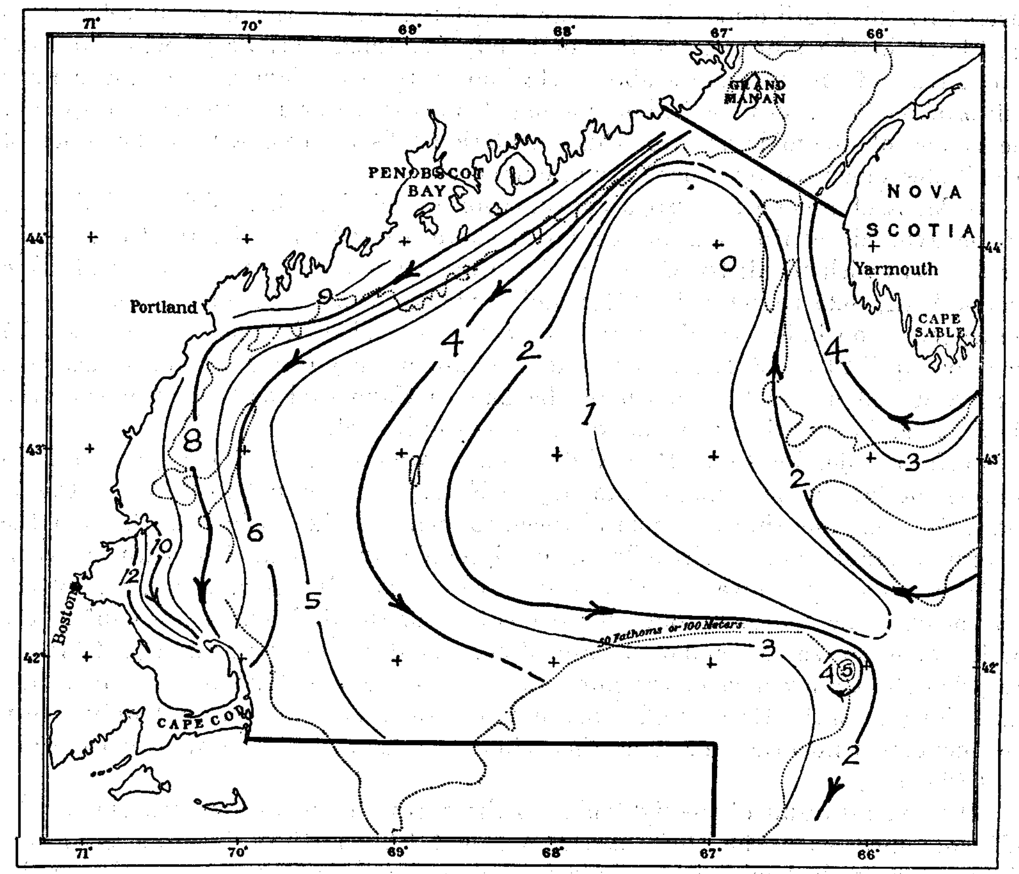

FIG. 192.—Dynamic gradient at the surface of the gulf, April 6 to 20, 1920, referred to the offing of the Bay of Fundy as base station. Contours for every dynamic centimeter

its offing (fig. 194) in a distance of 30-odd miles, but only about one-seventh as wide a difference at the 50 or 100 meter levels (stations 10266 and 10267).

As a result, the dynamic chart for May (fig. 195) corresponds closely to the distribution of density at the surface, except for the relationship between the shallows of German Bank and the deep water immediately to the west of the latter. In this region the surface projection, taken by itself, would give a false picture, being confused by the strong tides that keep the water thoroughly stirred over the bank, thus

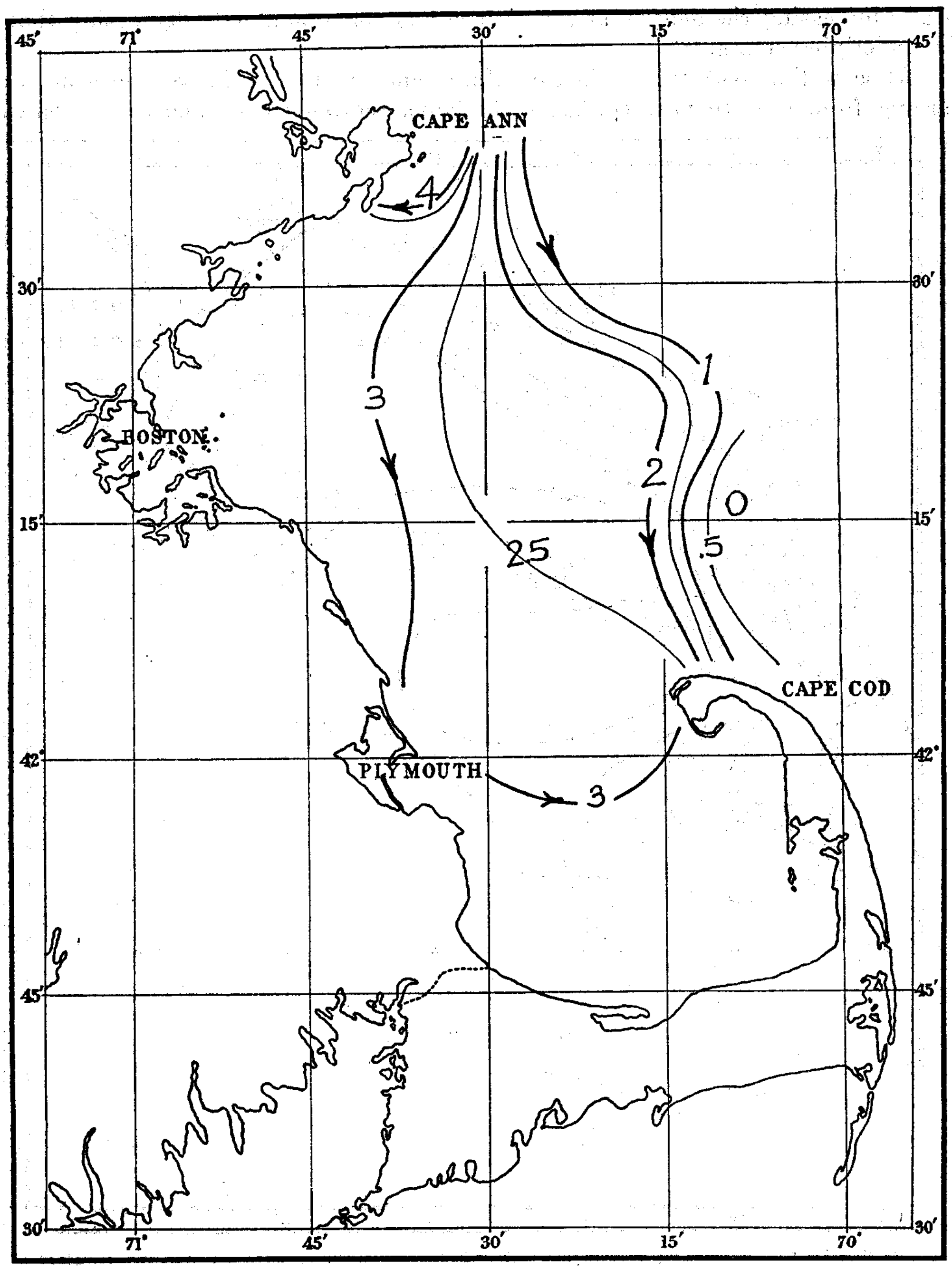

FIG. 193.—Dynamic gradient at the surface of Massachusetts Bay, April 21 to 23, 1925. Contours are for every one-half dynamic centimeter. Based on hydrometer readings

locally increasing the density at the surface but correspondingly decreasing that of the underlying strata.

At some time between the last of March and the first of May—the exact date varying from year to year (p. 832)—the Nova Scotian current, flooding westward

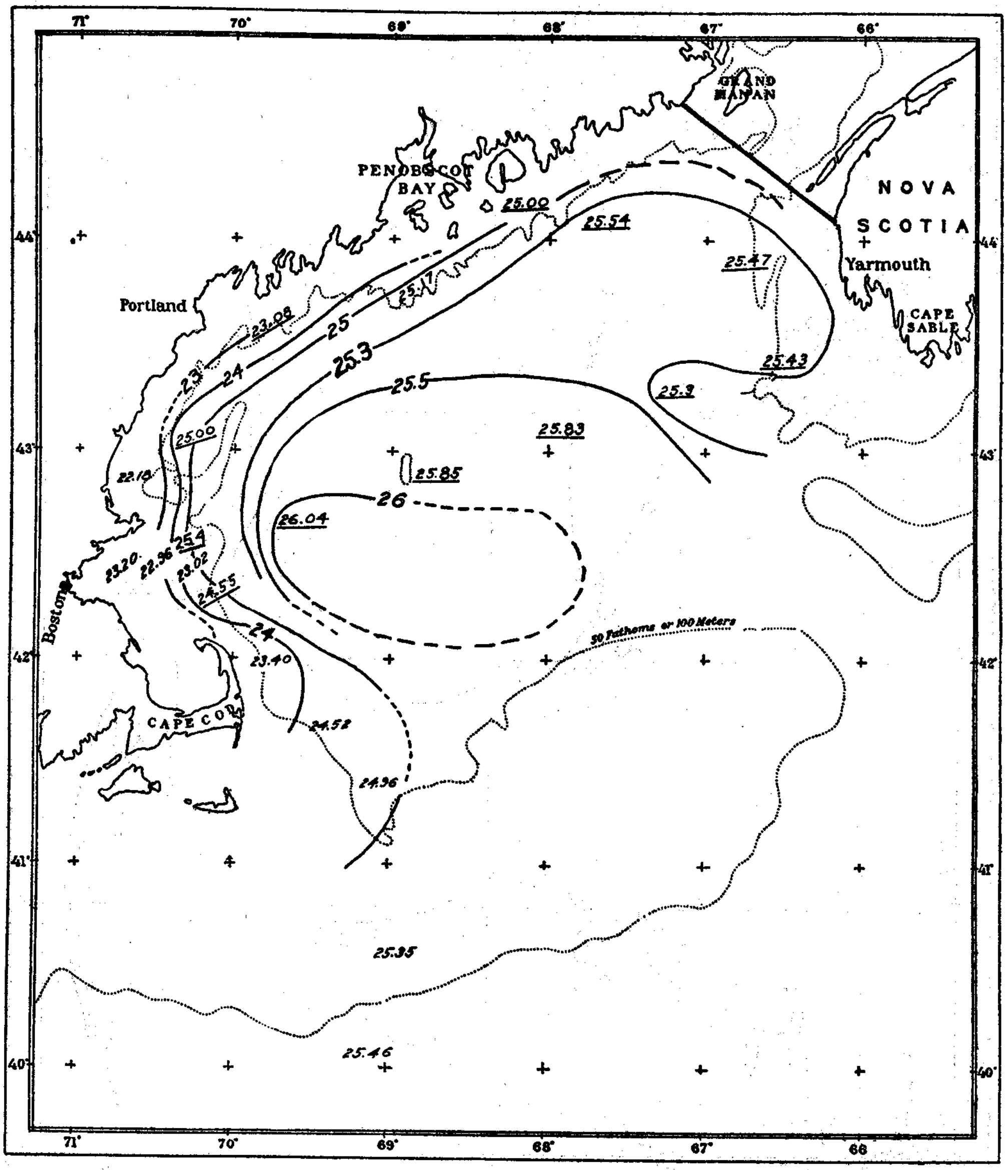

FIG. 194.—Distribution of density at the surface of the gulf or May, 1915 (underlined), and May, 1920, combined

past Cape Sable into the gulf, is reflected by the development of a corresponding tongue of low surface density extending westward from the offing of the cape. Thus, in 1919 the eastern half of the Cape Sable-Cape Cod profile proved less dense than the western in the upper 50 meters at the end of March and again at the end

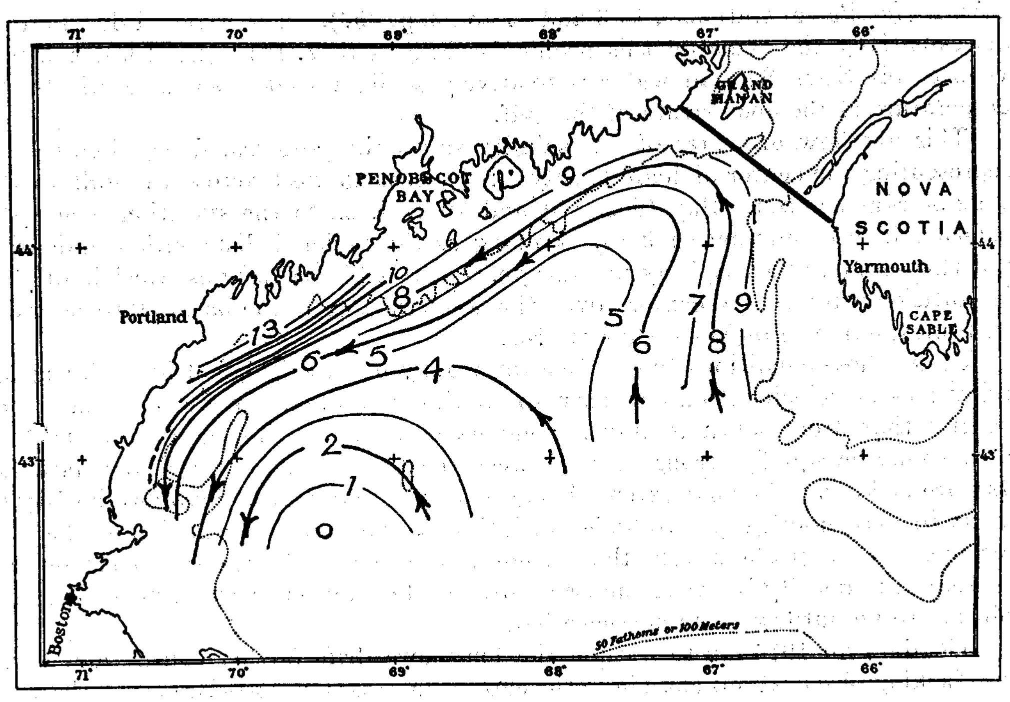

FIG. 195.—Dynamic gradient at the surface, for the northern part of the gulf, May 4 to 14, 1915, referred to the offing of Cape Ann as base station. Contour lines are for every dynamic centimeter

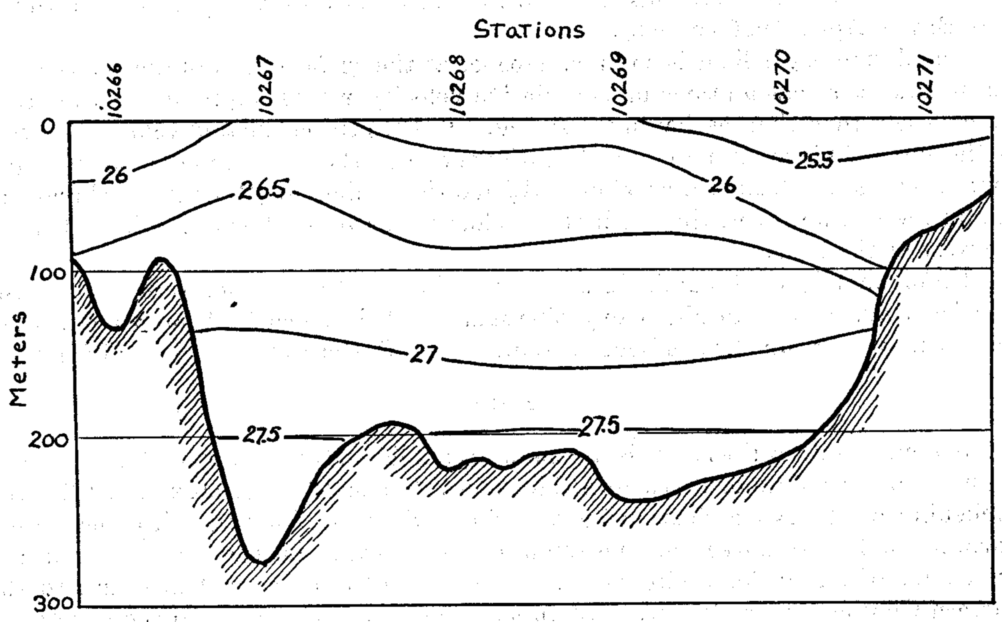

FIG. 196.—Density on a profile crossing the gulf from Massachusetts Bay toward Cape Sable, May 4 to 14, 1915. Corrected for compression

of April (Ice Patrol stations 2 to 3 and 21 to 23, p. 997). The regional distribution was essentially the same on this profile for May 4 to 7, 1915 (fig. 196), and it is because this Nova Scotian water is relatively so light that it so little affects the temperature of the deep strata of the gulf.

This overflow of water of low salinity shifts the potential depression, or low (representing the center of high density), from east to west across the gulf to the offing of Massachusetts Bay (figs. 194 and 195)—i. e., to the situation where the surface is high in summer (p. 956). So long as the regional distribution of density is of this sort (from early May in some years; probably as early as April in others) the anticlockwise vortex centers over the western arm of the basin 30 to 50 miles out from the mouth of Massachusetts Bay.

Under these conditions the surface water may be expected to drift with considerably greater velocity from northeast to southwest around the western margin of the gulf than from south to north along its eastern trough (fig. 195), though the current may be equally strong next the west coast of Nova Scotia, where data for May are lacking. To what extent this anticlockwise circulation involves the Bay of Fundy in that month is yet to be learned, though the sudden freshening of the surface there by the freshets from the St. John River (p. 808) suggests a considerable differential in density between the two sides of the bay as characteristic of May, pointing to an outflow in its northern half.

The data for 1915 fail to outline the longshore drift farther south than Cape Ann, lacking observations close in to the cape or in Massachusetts Bay, but the very low densities recorded at the mouth of the bay in May, 1920 (fig. 194), show it continuing down past Cape Cod, consistent with the drifts of bottles set out in Massachusetts Bay in April, 1926 (p. 893).

The dynamic gradient is so much steeper at the surface than in the deeps of the gulf in May that calculations of the relative velocity would approximate the truth more closely then than earlier in the spring. In 1915 the calculated velocity relative to the low off Cape Ann (assumed stationary, fig. 195) was about 0.23 knot per hour near Cape Elizabeth, or about 5½ nautical miles in 24 hours. Abreast of Mount Desert, however, the calculated velocity was only about 0.14 knot toward the west at the time.

Unfortunately no dynamic data are available for the southeastern part of the area for May, so that nothing can yet be said about the effect that the Nova Scotian current may exert on the gradient currents of the Eastern Channel and vicinity.

JUNE

No one of our cruises affords a general dynamic picture of the gulf as a whole in June, but the state of its eastern side shows that in 1915, at least (fig. 197), the slackening of the Nova Scotian current from the east, coupled with the vernal warming and progressive incorporation of land water in the west, caused the low center of anticyclonic circulation to shift from the offing of Cape Ann to the Eastern Channel by the last week of June. This seasonal return to the location it occupies in March (judging from 1920) probably represents the normal progression, the physical changes on which it depends being yearly events.

With this gradient a considerable indraft is indicated into the eastern side of the gulf; not, however, from the coastal belt to the eastward of Cape Sable, but from the region of Browns Bank and of its offing. Probably this indraft had as a counter current an outdraft from the gulf around the eastern end of Georges Bank, though, lacking a station on the bank, this can not be asserted definitely. It is certain, also, that the dynamic impulse for a northeast-southwest current around the northern and western margins of the gulf had slackened by the middle of that June.

Unfortunately, no observations were taken in the western side of the gulf that June, but a survey of Massachusetts Bay carried out by the *Fish Hawk* on June 16

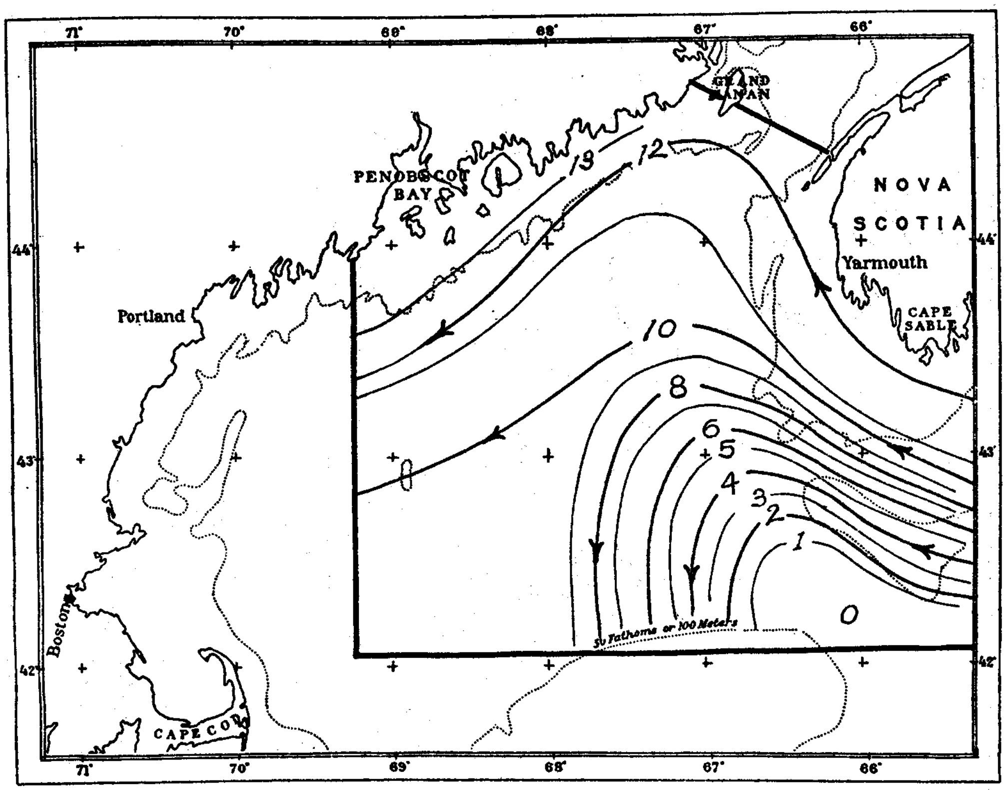

FIG. 197.—Dynamic gradient at the surface of the eastern side of the gulf, from June 10 to 26, 1915, referred to the Eastern Channel as base station. Curves are for every dynamic centimeter

and 17, 1925 (cruise 14), has enabled Mr. Parmenter to calculate the relative velocities and directions of the gradient current on various profiles by the method elaborated by Sandström (1919), and his results are offered here to illustrate this alternative procedure.

These calculations (tabulated below) rest on two assumptions—first, that the water was stationary at the greatest depth of the shoaler of each pair of stations, and, second, that the profiles selected (typical examples are shown in fig. 198) are at right angles to the existing current. In the present instance neither of these requirements is exactly fulfilled, but the close agreement between the calculation and the general distribution of density in the upper 20 meters (fig. 199) makes it probable that the calculated directions are a close approximation to the actual dynamic tendency toward circulation at the time.

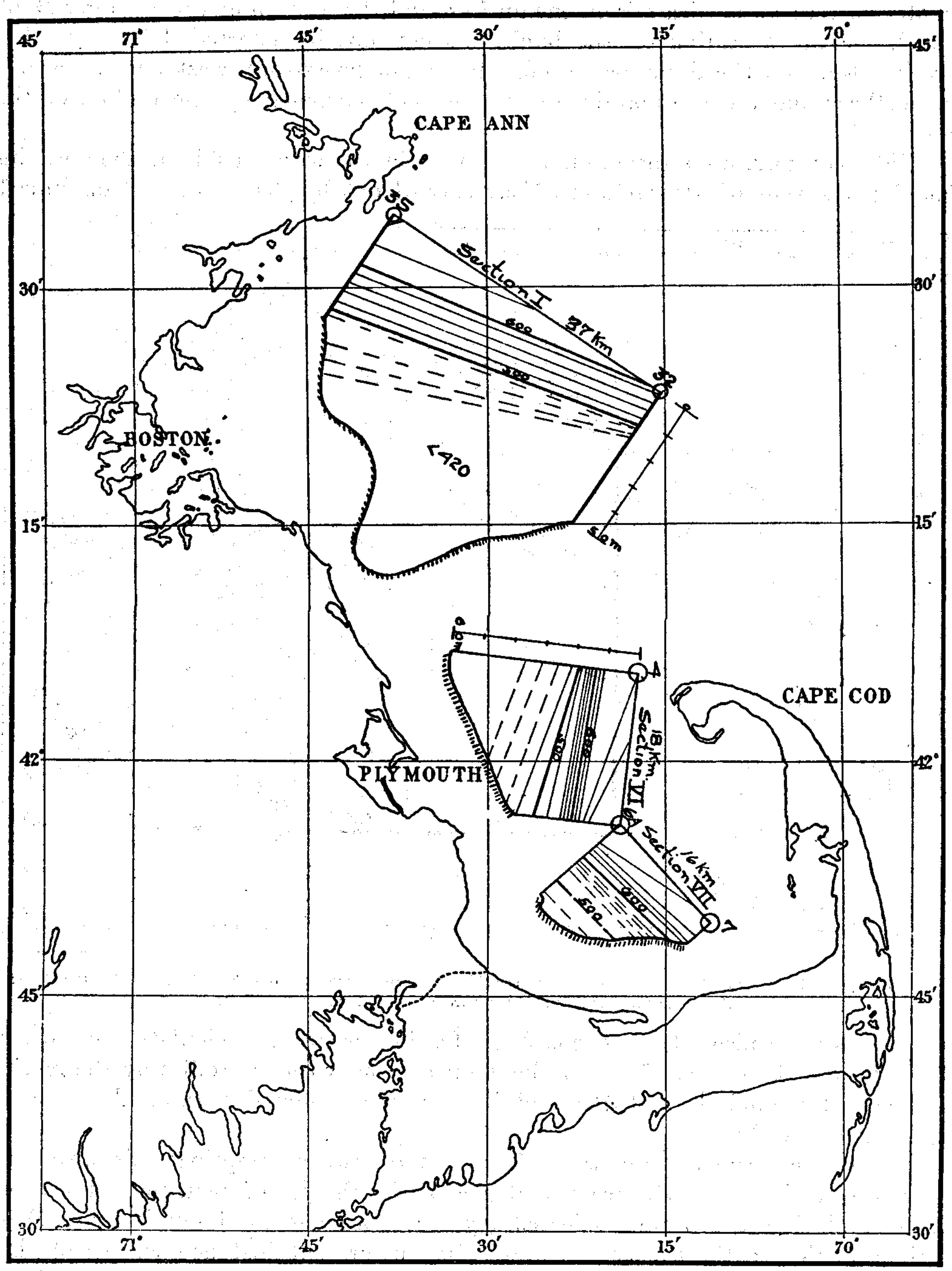

FIG. 198.—Specific volumes on three profiles in Massachusetts Bay, June 16 and 17, 1925. Calculated by R. Parmenter